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Studies on Diatoms of the Argentine Coast, the Drake Passage and the Bransfield Strait*

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Abstract: Forty-eight species of diatoms were identified from 37 stations occupied along the Argentine Continental Shelf, in the Drake Passage and in the Bransfield Strait in February-March, 1963. Of the species identified *Rhizosolenia alata*, *R. styliformis*, *Nitzschia seriata* had an ubiquitous distribution. In terms of relative abundance the number of diatoms collected at the stations occupied along the coast of Argentina was relatively poor compared to the Drake Passage and the Bransfield Strait stations. The highest number of diatoms count was found in the southern Drake Passage. The study of the composition of the diatoms in this cruise proved to be useful in the analysis of water masses and in delineating boundaries between them. This was especially true in the Drake Passage where variations in species composition and abundance of diatoms were used to corroborate the position of the Antarctic Convergence by the marked decrease in temperature.

1. Introduction

During the period February 15-March 20, 1963, the Argentine vessel ARA/*Capitan Canepa* occupied 48 stations along the Argentine Continental Shelf, in the Drake Passage and the Bransfield Strait (Figure 1). The main objectives of this cruise were to study primary production and standing crop of the phytoplankton in these areas and to correlate their distribution and abundance with the physical and chemical parameters studied (see EL-SAYED, 1963 and EL-SAYED *et al*, 1964). Since an appraisal of the productivity of an area would not be complete without a study of the species of phytoplankton which contribute to this production, efforts were made during this cruise to collect phytoplankton samples at the various stations occupied. Our main interest in this paper is to

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present an analysis of the species composition of the diatom population studied at 37 stations and to discuss their distribution and relative abundance in the areas investigated.

2. *Methods*

A Hensen-type "Kitahara" net (mesh size 0.09 mm) with a mouth opening of 24 cm in diameter was used for sampling the phytoplankton. A flow-meter was mounted at the mouth of the net to estimate the volume of the water filtered. Sampling was performed by towing the net vertically from a depth of 150 m to the surface; at some shallow stations off the Argentine Coast, hauls of less than 150 m were made. Since the flow-meter did not function properly at several stations, it was difficult to compare the volume of water strained at the sampling stations. Instead, the number of diatoms to be presented in this study will be expressed on a relative scale based on a 100 m vertical haul. Although the use of the net method for the quantitative assessment of phytoplankton is fraught with error, the underlying assumption in the interpretation of the data presented here is that since the same net was used throughout this cruise, and since the sampling technique (particularly speed of hauling) was executed in a similar manner at all the stations, net hauls can give a rough approximation of the phytoplankton population in the areas under investigation. The cell counts should, therefore, be regarded as representative of only the larger phytoplankters and not of the total species present. In other words, the numbers reported here would serve as a guide to the order of magnitude of the diatom concentration and should not be interpreted as absolute values.

All the phytoplankton samples were preserved with 10 % formalin. In the case of rich hauls, aliquots of the samples were placed on a counting plate (Rigosha Co. Ltd., Japan) and the numbers of cells were counted for each species; in the case of smaller samples, all the diatoms cells were counted.

3. *Results*

A total of 48 species of diatoms were identified in the samples collected during this cruise. Table 1 summarizes the pertinent information collected at the sampling stations. The same table also shows the percentage contribution of each species to the total diatoms population at each station. Shown in Figure 2 is the percentage distribution of the abundant diatoms collected at the stations

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occupied off the Argentine Coast, in the Drake Passage and in the Bransfield Strait during this cruise.

Of the 48 species of diatoms identified, *Rhizosolenia alata* BRIGHTWELL, *R. styliiformis* BRIGHTWELL, *R. styliiformis* var. *longispina* HUSTEDT and *Nitzschia seriata* CLEVE, had an ubiquitous distribution. They are noted for their wide geographic distribution from the tropics to high latitudes. Of these, *Nitzschia seriata* was particularly numerous in all the samples and was the dominant species in the southern part of the Drake Passage and along the northern coast of Argentina.

Dactyliosolen antarcticus CASTRACANE

Rhizosolenia alata fo. *inermis* (CAST.) HUSTEDT

R. bidens KARSTEN

Chaetoceros Castracanei KARSTEN

C. criophilus CASTRACANE

C. dictyota EHRENBERG

C. neglectus KARSTEN

Talassiothrix antarctica KARSTEN

Fragilaria antarctica CASTRACANE

The above nine species were observed mostly in Antarctic waters. The last two species were quite numerous in the Drake Passage.

Rhizosolenia alata fo. *gracillima* (CLEVE) GRUNOW

This species was observed along the northern coast of Argentina. It was the dominant species at Station 47 where it contributed about 80 per cent of the total cell count.

Rhizosolenia curvata ZACHARIAS

This species was observed mostly between the southern coast of Argentina and the Antarctic Convergence.

Rhizosolenia Shrubsolei CLEVE

This species was fairly numerous at Station 48. However, a few were also observed at Station 30 near the Antarctic Convergence.

Rhizosolenia simplex KARSTEN

Relatively few numbers were observed in the Drake Passage stations and at Station 10.

Chaetoceros affinis LAUDER

The distribution of this species was limited to the northern coast of Argentina.

Chaetoceros decipiens CLEVE

This species was quite abundant off the southern tip of Tierra del Fuego (Stations 16, 17 and 32) where it constituted nearly 50 per cent of the total diatom cells collected.

Thalassiothrix longissima CLEVE et GRUNOW

This species was observed along the southern coast of Argentina and in northern Drake Passage (Stations 18 and 19).

Eucampia balaustium CASTRACANE

This species was observed only in the Bransfield Strait.

Besides the diatoms given above, the following species were also identified in the samples collected during this cruise:

Coscinodiscus anguste-lineatus A. SCHMIDT

C. asteromphalus EHRENBERG

C. perforatus EHRENBERG

C. stellaris ROPER

Actinoptychus undulatus (BAIL.) RALFS

Asteromphalus regularis KARSTEN

Thalassiosira decipiens (GRUNOW)

Stephanopyxis turris (GREVILLE)

Dactyliosolen mediterraneus PERAGALLO

Rhizosolenia styliiformis var. *latissima* BRIGHTWELL

Chaetoceros convolutus CASTRACANE

C. densus CLEVE

C. pendulus KARSTEN

C. peruvianus BRIGHTWELL

C. debilis CLEVE

Triceratium arcticum BRIGHTWELL

Ditylum brightwellii GRUNOW

Eucampia zoodiacus EHRENBERG

Thalassionema nitzschioides HUSTEDT

Pleurosigma affine GRUNOW

Nitzschia delicatissima CLEVE

N. pungens var. *atlantica* GRAN

The number of diatoms in the 24 stations occupied along the coast of Argentina, in general, was relatively poor compared to the stations in the Drake Passage and the Bransfield Strait. The paucity of the diatom cells was conspicuously noted at Stations 5, 6 and 9 where less than 100 cells per 100 m haul were

encountered at each of these stations. Higher diatom counts, on the other hand, were found at Station 48 where more than 22,000 cells per unit haul were taken.

In the Drake Passage, the diatom population was exceptionally rich. At Station 32, south of Tierra del Fuego, the number of diatom cells (per unit haul) was in excess of 22,000 cells. A marked increase in number was noted at Stations 19 and 30, in the northern Drake Passage, where 72,000 and 148,800 cells were counted respectively. Stations 20, 29, 28 and 27, located to the south of the Antarctic Convergence, showed conspicuous increases in the number of cells counted; each station averaging well over a million cells per unit haul. Station 28, in the southern Drake Passage, where the number of diatom cells was more than 8 million was by far the richest station occupied during this cruise. It is interesting to point out in this respect that the distribution and relative abundance of the diatom cells in the Drake Passage found during this cruise was found to be quite similar to the observations made by the *Discovery* Expedition during February-March, 1930 (HART, 1934).

The number of diatom cells in the Bransfield Strait was unexpectedly low, ranging between 10^3 and 10^4 cells/unit haul, which is much lower than the number of cells encountered in the Drake Passage.

4. Discussion

Despite the shortcomings of using the net haul method in estimating the abundance of the phytoplankton population, it is instructive, nonetheless, to compare the results of this study with those based on the C^{14} uptake and the chlorophyll *a* methods taken during this cruise. Figure 3 shows that the cell count method (Fig. 3C) was in good agreement with the chlorophyll *a* content and primary production data (Fig. 3 A and B respectively) collected along the Argentine Coast and in the Drake Passage. It is difficult, however, to explain the disparity between the results of the two methods when applied to the Bransfield Strait stations, where EL-SAYED *et al* (*ibid.*) have shown higher pigment content and greater C^{14} uptake in the Bransfield Strait than in the Drake Passage. The reason behind this discrepancy is not clear. It is possible, however, that a good many of the phytoplankton organisms which contributed to the high primary production in the Bransfield Strait were composed of nanoplankton which escaped through the meshes of the net used. If so, this should point out to the significant role played by the nanoplankton in productivity studies.

The study of the diatom composition in this cruise proved to be useful in

the analysis of water masses and in delineating boundaries between them. For example the stations located south of Tierra del Fuego (Stations 16, 17 and 32) showed a closer similarity in their diatom composition than those to the east of Tierra del Fuego (Stations 33, 34 and 35) despite the geographical proximity of the two groups of stations. In the former group a similar diatom composition was noted with the dominant species being: *Nitzschia seriata*, *Rhizosolenia styliiformis*, *Chaetoceros atlanticus* and *C. decipiens*; the latter species made up about one half of the total cell number collected at each of these stations. At Stations 33, 34 and 35, on the other hand, the dominant species were: *Rhizosolenia alata*, *R. styliiformis* and *Chaetoceros atlanticus*, each contributing about the same percentage to the total number of cells collected. On the basis of this diatom association, it is possible to postulate that the station located to the east of Tierra del Fuego, and those to the south of it are under the influence of two different water masses.

Another interesting finding which emerged from our study of the species composition of the diatoms was that at Station 48, in the northern Argentine Coast, Antarctic species such as *Corethron criophilum*, *Dactyliosolen antarcticus*, *Chaetoceros criophilus* and *C. dichaeta*, which were hardly seen at other coastal stations, were observed at this station. The fact that these Antarctic species were observed at this station may well indicate isolated eddies from the Antarctic Ocean even to this distant locality. Similar observations were also found by BALECH (1959) where at two stations located as far north as 41°26'S some Antarctic plankters were found. These included among the dinoflagellates, *Peridinium parvicollum* and *P. turbinatum*. Among the diatoms* there were: *Asteromphalus hookeri*, *Chaetoceros criophilus*, *Coscinodiscus antarcticus*, *Fragilariopsis antarctica*, *Fragilariopsis linearis curta*, *Synedra reinboldi*. Of the less typically Antarctic species: *Chaetoceros atlanticus*, *Corethron criophilum*, *Dactyliosolen antarcticus* and *Eucampia balaustium* were encountered at the two aforementioned stations.

In the Drake Passage, the two northern stations (18 and 19) had a very similar diatom population which was quite different from Station 30 in the middle of the Passage. At the former stations, *Rhizosolenia alata* contributed more than 70 per cent of the total diatom count with *Rhizosolenia curvata*, *R. simplex* and *Thalassiothrix longissima* also present. At Station 30, on the other hand, a marked floral change took place, the dominant species being *Dactyliosolen*

* Identified by FRANGUELLI (BALECH, personal communication).

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antarcticus, *Rhizosolenia alata* fo. *inermis*, *R. bidens*, *Chaetoceros dichaeta* and *Thalassiothrix antarctica*. South of this station to the South Shetland Islands, the diatom composition was very similar throughout that region with *Rhizosolenia alata*, *Thalassiothrix antarctica*, *Chaetoceros atlanticus* and *Nitzschia seriata* predominating. On the basis of the diatom composition in the Drake Passage, it can be reasonably assumed that the Antarctic Convergence is located in the vicinity of Station 30. This is in good agreement with the data presented by EL-SAYED *et al* (*ibid.*) who used the chlorophyll *a* and C¹⁴ uptake data to corroborate the position of the Antarctic Convergence by the marked decrease in temperature (from 5.0°C to 2.2°C) between Station 31 and 30.

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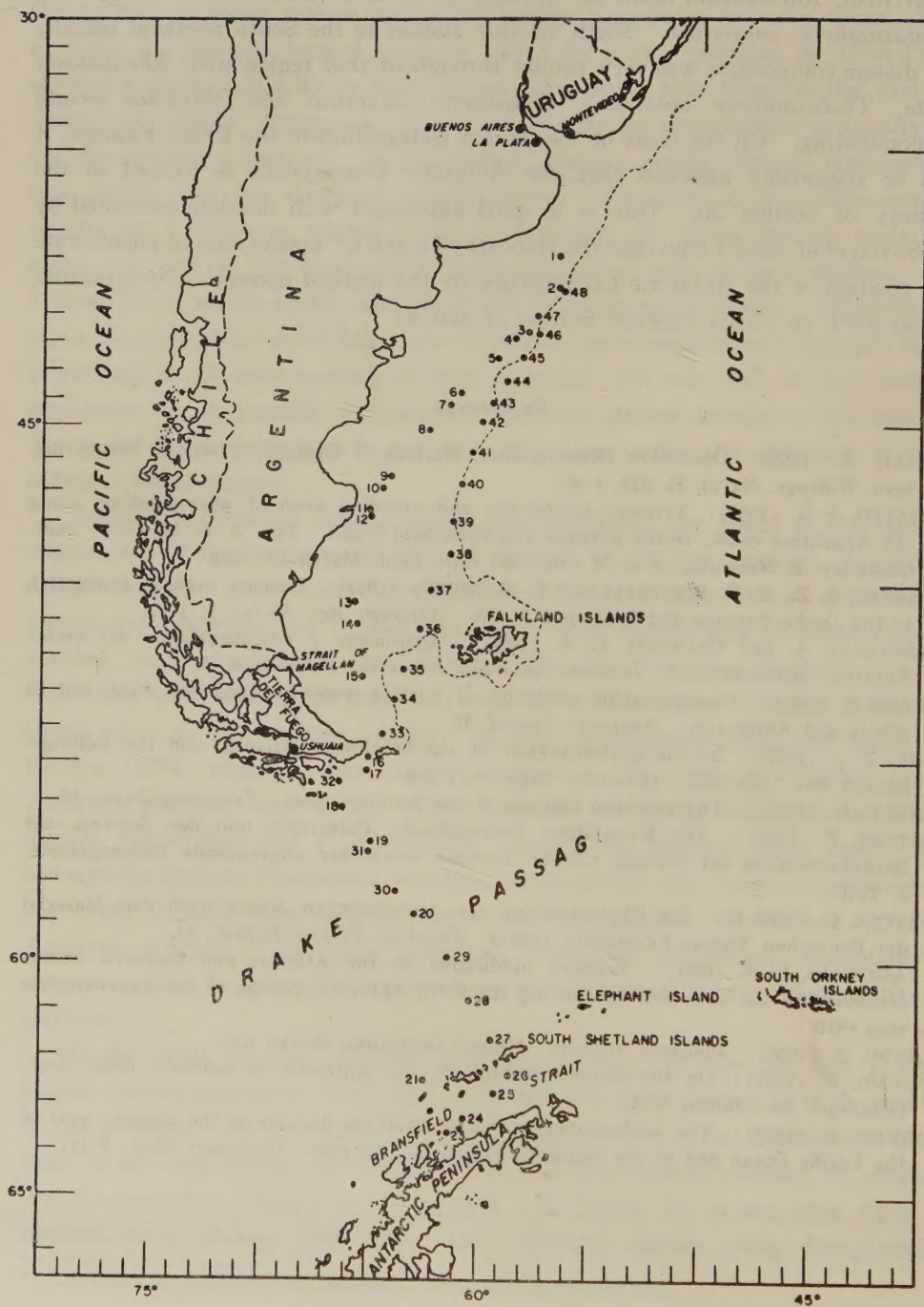


Fig. 1. Position of stations occupied along the Argentine Continental Shelf, in the Drake Passage and in the Bransfield Strait. (February-March, 1963)

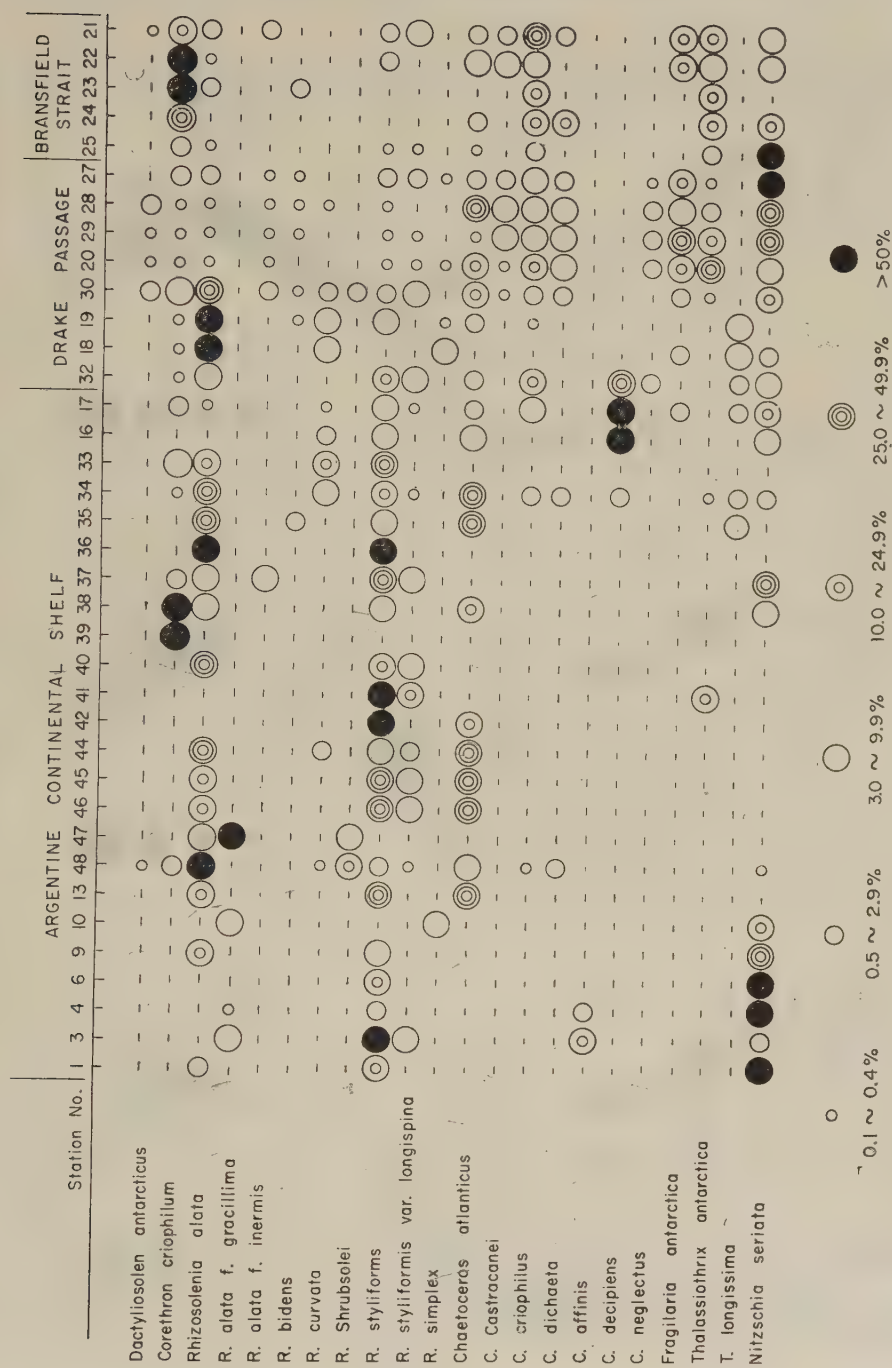


Fig. 2. Percentage distribution of diatoms collected at the sampling stations along the Argentine Continental Shelf in the Drake Passage and Bransfield Strait. (February-March, 1963)

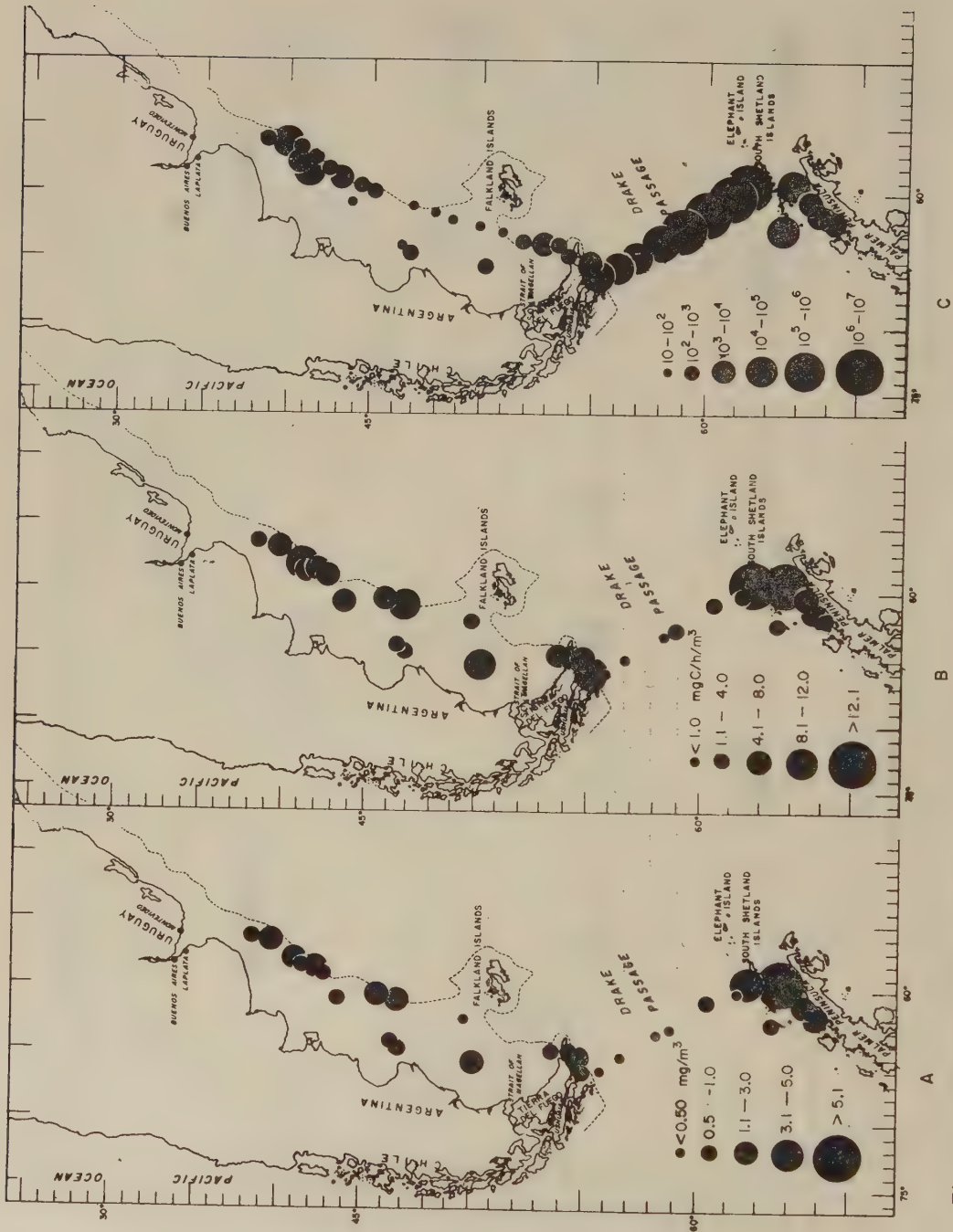


Fig. 3. Distribution of Chlorophyll *a* (A), C^{14} uptake (B) and number of diatoms per 100 m vertical haul (C), taken along the Argentine Continental Shelf, in the Drake Passage and Bransfield Strait. (February-March, 1963)

Mean Sea Level in Bays, with Special Reference to the Mean Slope of Sea Surface due to the Standing Oscillation of Tide

By

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Abstract: The distribution of mean sea level around the Japanese Islands was outlined in the beginning (Fig. 3), and it was clearly recognized that the sea level is high on the coast located on the right of ocean currents and low on the coast located on the left of them. Especially, it was pointed out in many of the bays that the mean sea level is higher in the inside of the bay than in the vicinity of the entrance of it (Fig. 5). To interpret this fact, the authors investigated effects of several factors such as bottom friction, meteorological elements, ununiform distribution of density of sea water *et cetera* which might have influence on the mean sea level in the bay, but satisfactory result was not gained. At last the mean slope of sea surface of the second order due to the standing oscillation of tide was considered. And, from the result of numerical computations, it was concluded that the majority of conspicuous rise of mean sea level found in both Bays of Ariake and Ise is attributed to the tidal oscillation. As for other bays the circumstances remain uncertain, although it appears in some cases that the amount of rise of mean sea level is within errors contained in the materials.

1. Mean sea level on the coast of Japan

The rough distribution of mean sea level on the coast in our country will be sketched first, and then we will investigate the mean slope of sea surface in bays in detail. According to the report presented in 1925 by the Land Survey Department (now renamed to the Geographical Survey Institute), it was clearly recognized that the mean sea surface slopes along the coast of Japanese Islands (Land Survey Department, 1925). Recently the same authorities reexamined it (Geographical Survey Institute, 1960). The main results in both cases are shown in Table 1. As a reference level we use the Standard Datum Surface of the Japanese Leveling Network. It is usually called as the Mean Sea Level of Tokyo Bay (Tokyo-wan chuto-choi in Japanese) or Tokyo Point (abbr. T. P.), because it was

Table 1. Mean sea level, referred to the Standard Datum Surface of the Japanese Levelling Network (Tokyo wan chuto-choi in Japanese or T. P.), at tidal stations attached to the Land Survey Department or the Geographical Survey Institute. The latitude and longitude of the stations are given in Table 2. After Land Survey Department (1925) and Geographical Survey Institute (1960).

Result presented in 1925					Result presented in 1960			
Tidal station	Mean sea level	Period of calculation of M.S.L.	Start of recording	Epoch of levelling	Tidal station	Mean sea level	Period of calculation of M.S.L.	Period of levelling
	cm	years				cm	+1900	+1900
Aburatsubo	+ 0.3	23	1892		Aburatsubo	- 4.0	41-59 ⁽²⁾	49-52
Ayukawa	+ 1.5	24	1892	1900	Kainan	+ 6.2	54-59	47-52
Kushimoto	+10.8	24	1892	1899	Kochi	- 9.7	51-59	47-52
Hosojima	+11.1	24	1892	1894	Hosojima	+13.8 ⁽¹⁾	41-59	47-58
Fukabori	+22.4	24	1892	1895	Wajima	+24.2	41-59	48-52
Tonoura	+21.4	24	1892	1892	Kashiwazaki	+31.4	56-59	50-56
Wajima	+24.7	24	1892	1900	Nezugaseki	+35.6	56-59	50-56
Iwasaki	+25.7	24	1893	1903	Oshoro	+17.2	41-59 ⁽³⁾	50-59
Oshoro	+ 2.9	24	1902	1905	Asamushi	+21.0	55-59	50-59
Hanasaki	- 0.8	24	1896	1903				

(¹) temporary result, (²) except 1945-47, (³) except 1946-47.

originally referred to the mean sea level observed at Reigan-jima in Tokyo for six years and a half from 1873 to 1879. The values given in Table 1 seem to be comparatively reliable on the whole because careful attention and great effort were paid to connect the elevation of the "Fixed Point" at each tidal station with T. P. An interesting map showing the general distribution of water level on the coast of our country was also presented by the Geographical Survey Institute (1960), but a little blank is recognized in it on account of scantiness of data.

So, to fill up the blank, we collected the published data of mean sea level that was capable of being connected with T. P. The result is arranged in Table 2 as well as the source of the data. The localities of the stations are shown in Fig. 1 with numerals given in Table 2 in place of respective names. As the period of calculation of mean sea level, these five years were used in most cases to increase the number of materials and to make all of uniform

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Table 2. Mean sea level, referred to the Standard Datum Surface of the Japanese Levelling Network (Tokyo wan chuto-choi in Japanese or T. P.), at tidal stations in Japan. The localities of the stations are shown in Fig. 1.

No. in Fig. 1	Tidal Station	North Lat.	East Long.	M.S.L. above Station Datum Level	Height of T.P. above Station Datum Level	Mean Sea Level above T.P.	Period of Calcula- tion of M.S.L.	Source* of Date	Year** of Levell- ing
		° ' "	° ' "	cm	cm	cm	+1900		
1	Hanasaki	43 17	145 35	213.6	200.7	+12.9	59-63	M	1955
2	Kushiro	42 58	144 22	147.6	139.2	+ 8.4	59-63	M	1955
3	Urakawa	42 10	142 46	96.2	96.5	- 0.3	—	H	
4	Hakodate I	41 47	140 44	145.0	135.4	+ 9.6	59-63	M	1955
5	Hakodate II	41 46	140 43	127.4	154.3	-26.9	58-61	H	
6	Aomori	40 50	140 46	99.1	70.2	+28.9	59-63	M	
"	"	"	"	99.1	86.0	+13.1	59-63	S	
7	Asamushi	40 54	140 52	-291.9	-311.7	+19.8	59-63	G	
8	Hachinohe	40 32	141 31	72.0	50.3	+21.7	59-63	M	1957
9	Miyako	39 38	141 58	117.6	100.7	+16.9	59-63	M	1957
10	Kamaishi	39 16	141 54	225.6	222.4	+ 3.2	57-61	H	
11	Ayukawa	38 18	141 31	168.9	154.4	+14.5	59-63	M	1933
12	Shiogama	38 18	141 02	93.0	108.0	-15.0	—	S	
13	Onahama	36 56	140 55	145.2	135.5	+ 9.7	59-63	M	1956
"	"	"	"	145.2	133.5	+11.7	59-63	S	
14	Choshi	35 44	140 50	104.3	110.7	- 6.4	59-63	M	1953
15	Mera	34 55	139 50	122.6	124.6	- 2.0	59-63	M	
16	Tokyo	35 39	139 46	183.8	180.0	+ 3.8	59-63	M	1958
17	Kawasaki	35 30	139 45	152.2	137.1	+15.1	59-63	M	1953
18	Yokohama	35 29	139 41	177.0	174.7	+ 2.3	59-61	M	
"	"	"	"	177.0	168.8	+ 8.2	59-61	S	
19	Yokosuka	35 17	139 39	150.3	142.6	+ 7.7	57-61	H	
"	"	"	"	110.0	108.3	+ 1.7	—	S	
20	Aburatsubo	35 09	139 37	-320.8	-320.8	0.0	59-63	G	
21	Ito	34 58	139 06	209.9	213.6	- 3.7	59-63	M	1953
22	Uchiura	35 01	138 54	163.1	165.2	- 2.1	59-63	M	1953
23	Shimizu-ko	35 00	138 30	146.5	142.2	+ 4.3	59-63	M	1953
24	Omaezaki	34 36	138 14	159.1	167.9	- 8.8	59-63	M	1953
25	Maisaka	34 41	137 37	56.1	78.4	-22.3	59-63	M	1953
26	Nagoya	35 05	136 53	208.7	202.5	+ 6.2	59-63	M	1953
27	Toba	34 29	136 51	263.2	268.6	- 5.4	59-63	M	
28	Owase	34 04	136 12	158.2	173.9	-15.7	59-63	M	1953
29	Uragami	33 33	134 54	145.2	168.8	-23.6	59-63	M	1953
30	Kushimoto	33 28	135 46	167.4	182.6	-15.2	59-63	M	1953
"	"	"	"	170.0	182.8	-12.8	55-59	K	1953
31	Shimotsu	34 06	135 09	192.0	188.2	+ 3.8	59-63	M	1953
"	"	"	"	195.7	188.6	+ 7.1	55-59	K	1953
32	Kainan	34 09	135 12	-264.3	-267.9	+ 3.6	59-63	G	
33	Wakayama	34 13	135 09	103.4	102.2	+ 1.2	59-63	M	1953
"	"	"	"	109.7	102.2	+ 7.5	55-59	K	1953
34	Tannowa	34 20	135 10	186.0	180.2	+ 5.8	59-63	M	1953
"	"	"	"	189.2	180.5	+ 8.7	55-59	K	1953
35	Osaka	34 39	135 26	324.6	317.4	+ 7.2	63-64	M	1963
"	"	"	"	287.3	265.2	+22.1	59-60	K	1955
36	Kobe	34 41	135 11	136.7	130.9	+ 5.8	59-63	M	1953
"	"	"	"	129.5	114.9	+14.6	55-59	K	1953
37	Sumoto	34 20	135 54	184.0	167.8	+16.2	59-63	M	1953
"	"	"	"	185.2	169.3	+15.9	56-59	K	1953
38	Uno	34 29	133 57	165.5	140.4	+25.1	59-63	M	1953
"	"	"	"	166.1	146.4	+19.7	55-59	K	1953

No. in Fig. 1	Tidal Station	North Lat.	East Long.	M.S.L. above Station Datum Level	Height of T.P. above Station Datum Level	Mean Sea- Level above T.P.	Period of Calcula- tion of M.S.L.	Source* of Data	Year** of Levelling
		° /	° /	cm	cm	cm	+1900		
39	Kure I	34 14	132 33	257.8	235.3	+22.5	59-63	M	1953
"	"	" "	" "	261.7	235.2	+26.5	55-59	K	1953
40	Kure II	34 14	132 32	260.0	234.5	+25.5	57-61	H	
41	Tokuyama	34 02	131 48	256.3	243.5	+12.8	57-61	H	
42	Komatsujima	33 59	134 37	193.2	180.4	+12.8	59-63	M	1953
"	"	" "	" "	194.7	180.7	+14.0	55-59	K	1953
43	Takamatsu	34 21	134 03	190.5	170.8	+19.7	59-63	M	1953
"	"	" "	" "	190.6	172.8	+17.8	55-59	K	1953
44	Matsuyama	33 52	132 43	228.1	228.4	- 0.3	59-63	M	1953
"	"	" "	" "	232.4	224.8	+ 7.6	55-59	K	1953
45	Uwajima	33 14	132 33	218.3	216.8	+ 1.5	59-63	M	1953
"	"	" "	" "	220.0	216.5	+ 3.5	55-59	K	1953
46	Tosashimizu	32 47	132 58	167.8	180.2	-12.4	59-63	M	1953
"	"	" "	" "	169.1	179.8	-10.7	55-59	K	1953
47	Kochi I	33 30	133 35	113.9	131.5	-17.6	59-63	M	1953
"	"	" "	" "	118.8	131.0	-12.2	55-59	K	1953
48	Kochi II	33 30	133 34	-291.9	-277.6	-14.3	59-63	G	
49	Hosojima	32 26	131 40	-262.6	-267.0	+ 4.4	59-63	G	
50	Aburatsu	31 35	131 25	148.8	140.9	+ 7.9	59-63	M	
51	Kagoshima	31 36	130 34	200.6	191.3	+ 9.3	59-63	M	1946
52	Makurazaki	31 16	130 18	268.2	269.7	- 1.5	59-63	M	
53	Misumi	32 37	130 27	418.8	383.6	+35.2	59-63	M	
54	Miike	33 00	130 23	—	—	+32.0	—	N	
55	Oura	32 59	130 13	318.4	280.0	+38.4	51-52	N	
56	Fukabori	32 41	129 49	251.6	250.9	+ 0.7	59-63	M	1930
57	Megami	32 43	129 51	247.5	238.4	+ 9.1	50-60	M	1930
58	Sasebo	33 10	129 43	276.5	256.6	+19.9	57-61	H	
59	Moji	33 57	130 58	289.7	274.1	+15.6	58-61	H	
60	Shimonoseki	33 58	130 57	253.5	245.4	+ 8.1	59-63	M	1953
61	Tonoura	34 54	132 04	101.5	83.3	+18.2	59-63	M	1953
62	Sakai	35 33	133 14	130.9	114.3	+16.6	59-63	M	1953
63	Saigo	36 12	133 20	100.2	111.1	-10.9	59-63	M	1953
64	Miyazu	35 32	135 12	107.0	86.5	+20.5	59-63	M	1953
65	Maizuru	35 29	135 29	112.9	92.9	+20.0	57-61	H	
66	Wajima	37 24	136 54	-192.9	-214.6	+21.7	59-63	G	
67	Kashiwazaki	37 21	138 31	-252.6	-284.0	+31.4	59-63	G	
68	Nezugaseki	38 34	139 33	-210.2	-243.8	+33.6	59-63	G	
69	Iwasaki	40 35	139 54	165.6	129.2	+36.4	59-63	M	1957
70	Oshoro	43 12	140 52	-163.8	-177.8	+14.0	59-63	G	
71	Wakkanai	45 25	141 41	167.1	148.0	+19.1	59-63	M	1949
72	Monbetsu	44 21	143 22	348.9	337.2	+11.7	57-61	H	
73	Abashiri	44 01	144 17	145.7	141.7	+ 4.0	59-63	M	1955

* Source of Data

G : Geographical Survey Institute; Tidal Record, 1959-1963.

H : Hydrographic Office; Hydrographical Bulletin, 1964, No. 77.

K : Kobe Marine Observatory; Technical Report of Japan Meteor. Agency, 1962, No. 12.

M : Japan Meteorological Agency; Tide Tables for the year 1965.

N : Nagasaki Marine Observatory; see TERADA *et al* (1954).

S : Second Harbor Construction Bureau; Tech Rep. Y-094, 1965.

** From "The list of reference level at tidal stations" by Japan Meteorological Agency (1964). This column indicates the year when the bench mark of the first order in the vicinity of respective tidal stations was connected with the Standard Datum Surface by the first order precise levelling executed by the Geographical Survey Institute.

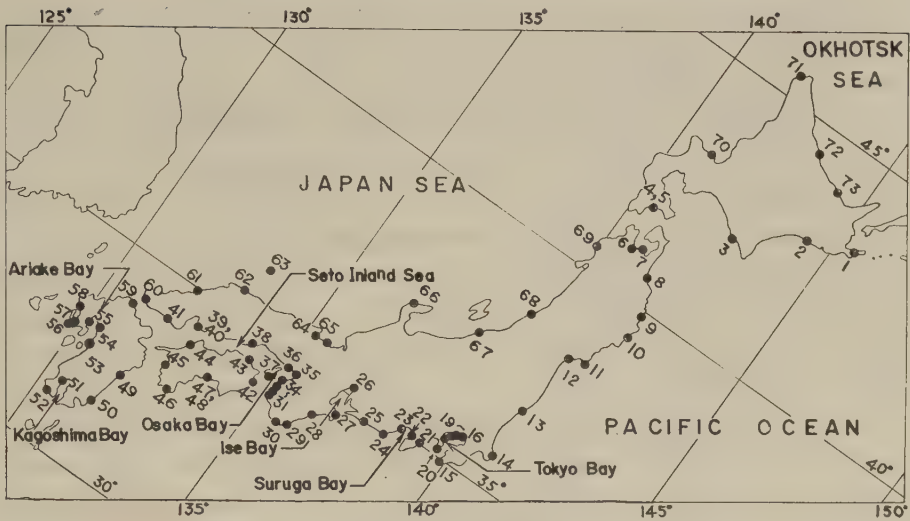


Fig. 1. Localities of tidal stations. Their names corresponding to the numerals are given in Table 2.

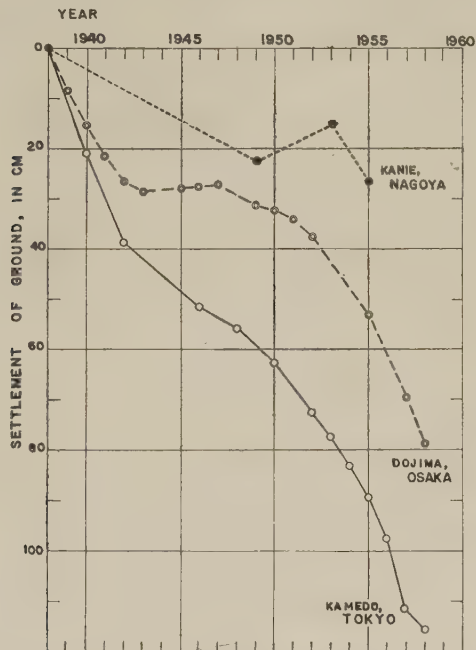


Fig. 2. Settlement of ground at three cities, Tokyo, Osaka and Nagoya. (The figure was made based on the materials of the Geographical Survey Institute.)

condition as much as possible. At many stations, however, rather great discrepancies are found between the period for averaging the tide level and that of precise levelling by which the datum level at each tidal station could be connected with T. P. Therefore, a fairly large error is likely to arise in regions where the movement or settlement of ground is active. Such regions are not rare in our country, particularly the settlement is noticeable in many industrial regions located around the innermost part of the bay as plainly seen in Fig. 2. Moreover, the error of levelling is also inevitable, and the longer the distance of levelling, the more the error. It is said that the mean of this error is about ± 10 mm near Tokyo, ± 20 mm in Chubu District, ± 30 mm in Kinki District, $\pm 40 \sim 45$ mm in Chugoku and Shikoku Districts as well as near Kushimoto, and ± 60 mm or less in Kyushu and Hokkaido (Geographical Survey Institute, 1960). On the other hand, since the mean sea level itself fluctuates depending on oceanographical, meteorological and astronomical conditions, a level for a certain period is different from a level for another period. Many examples in Tables 1 and 2 showing disagreement between mean sea levels for different periods at the same station are partly attributed to this, while some of them present apparent fluctuation due to the movement or the settlement of the ground and small discrepancy may come from the errors.

Thus it must be kept in mind that the values presented in Table 2 possibly contain considerable errors with $\pm$ several centimeters or more in some cases, and they also fluctuate according to the period of averaging. In order to diminish such errors, it is hoped that the datum line at every tidal station in our country is connected rapidly and simultaneously with T. P. by precise levelling, although the execution seems to be very difficult on account of enormous labor and expense being demanded.

Taken all together from the table, however, we can recognize a systematic distribution of mean sea level in spite of such fault being found in respective values. On the basis of Tables 1 and 2 we drew up Fig. 3 which roughly shows the recent distribution of mean sea level above T. P., considering the system of ocean currents in the vicinity of Japan. The typical paths of main currents are also added in the same figure. It is seen that the mean sea level is very high along the Japan Sea coast from Kyushu to Hokkaido, especially with the height more than 30 cm above T. P. on the northwestern coast of Honshu. The maximum is about 36 cm at Nezugaseki and Iwasaki. Along the east coast of Honshu the sea surface declines gradually from the north to the south, and the



Fig. 3. Rough distribution of mean sea level along the coast of the Japanese Islands, referred to the Standard Datum Surface of the Japanese Levelling Network (or T.P.). The means were taken over recent five years in most cases.

low level below T.P. covers the Pacific coast from Boso Peninsula to Kyushu. Particularly on the southeastern coast of Kii Peninsula the sea surface is the lowest with the level a little lower than -15 cm. Along the Okhotsk Sea coast of Hokkaido the sea level descends gradually towards the southeast, while on the Pacific coast the sea level is not high. It is noticeable that pretty large difference in sea level is found between both sides in Tsugaru Straits where rather conspicuous Tsugaru Current flows. Along the East Sea coast of Kyushu the sea surface seems to ascend from the south to the north.

Comparing with the system of ocean currents we can easily find in general that the mean sea level is high on the coast located on the right of the current and low on the coast located on the left of it. This suggests the effect of Coriolis force due to the rotation of the earth. But the detailed discussions will be left in future. (We see remarkable difference of the level between Sakai and Saigo situated on both sides of Tsushima current, but the value at Saigo in the Oki Islands is considerably unreliable naturally.)

2. Mean sea level in bays

Fig. 4, which we owe to the Nagasaki Marine Observatory, indicates the mean

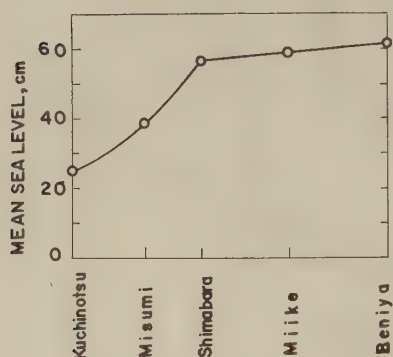


Fig. 4. Slope of mean sea level, taken over Oct. 1-15, 1953, in Ariake Bay. See Fig. 10 as to the localities of the stations. After TERADA, YASUI, AKAMATSU and NAKAMURA (1954).

slope of sea surface in Ariake Bay where the most predominant tide in our country arises with the range of five meters or so (TERADA, YASUI, AKAMATSU and NAKAMURA, 1954). The data are based on the tidal observation for half a month and levelling done at the same period. It is interesting that the surface rises abruptly about 30 cm from the mouth of the bay to the central region of it, but ascent becomes gradual towards the bay bottom.

As regards other bays the mean sea level above T.P. is shown in Fig. 5 based on the data of Table 2. In Tokyo Bay, the mean sea level is higher a little inside than outside, but the difference of height seems not to be far beyond the possible errors except a doubtful value at Kawasaki. At each station on the coast of Osaka Bay and Kii Channel, we find two values due to different sources shown in Table 2. Though somewhat suspicious values are found at a few stations, we could not deny the fact that the sea level in the arm of the sea is higher in comparison with one on the open sea. In Ise Bay too, the sea level at the inner end of the bay (Nagoya) is about 10 cm higher than the level at the entrance of it (Toba). It is noteworthy that the interior of Ariake Bay belongs to one of the sea areas with the highest mean sea level in our country, as well as the northwestern coast of Honshu (Fig. 3). At Nagasaki Harbor to the west

S. Unoki and I. Isozaki: Mean Sea Level in Bays

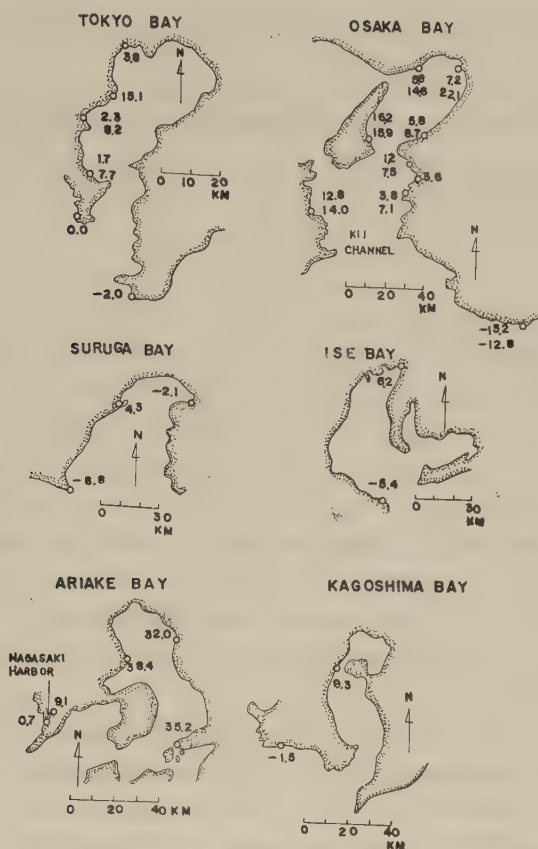


Fig. 5. Mean sea level in several bays, based on the data in Table 2 (unit, cm). See Fig. 1 for the localities of the bays.

of Ariake Bay, the sea level at Megami situated in the middle of this slender basin is about 8 cm higher than the sea level at Fukabori near the entrance. Similar situations are recognized in both deep bays of Kagoshima and Suruga, the greatest depth of water exceeding 200 m in the former and 1500 m in the latter. Moreover, the mean sea level in the Seto Inland Sea is pretty high as clearly recognized in Fig. 3.

Thus, so far as the data in Table 2 are based on, we may roughly say that the mean sea level near the bay bottom is more or less higher than that near the bay mouth in general, though it must be again remembered that considerable errors are likely to be contained in the present data as stated in the previous section.

3. *Examination of possible factors which might influence the slope of mean sea level in a bay*

When the hydrostatic pressure is assumed, the equations of motion are

$$\frac{\partial u}{\partial t} + u \frac{\partial u}{\partial x} + v \frac{\partial u}{\partial y} - f v = -g \frac{\partial \zeta}{\partial x} - \frac{1}{\rho_w} \frac{\partial p_s}{\partial x} + \frac{\tau_s^{(x)}}{\rho_w D} - \frac{\tau_b^{(x)}}{\rho_w D} \quad (1)$$

$$\frac{\partial v}{\partial t} + u \frac{\partial v}{\partial x} + v \frac{\partial v}{\partial y} + f u = -g \frac{\partial \zeta}{\partial y} - \frac{1}{\rho_w} \frac{\partial p_s}{\partial y} + \frac{\tau_s^{(y)}}{\rho_w D} - \frac{\tau_b^{(y)}}{\rho_w D}, \quad (2)$$

where x and y are horizontal rectangular coordinates taken on the undisturbed sea surface, u and v the components of mean velocity with depth, f the Coriolis parameter, ρ_w the density of sea water, g the acceleration of gravity, ζ the elevation of sea surface, p_s the atmospheric pressure on the sea surface, $\tau_s^{(x)}$ and $\tau_s^{(y)}$ the components of surface stress due to a wind, and $\tau_b^{(x)}$ and $\tau_b^{(y)}$ the components of bottom friction. When the depth of a basin is presented by h , the momentarily changing depth D is given by $h + \zeta$, but we may replace it by h alone in usual case. Strictly speaking, the nonlinear terms in the equations must be multiplied by suitable factors relating to the distribution of the current with depth, but we didn't add them for simplicity because they are nearly unity.

A. *Effect of bottom friction*

In a bay where a constant current is recognized, the sea surface slopes so that it may flow against the bottom friction. In the case of one dimension, the difference in sea level $\Delta \zeta$ between the points at a distance of Δl is given by

$$\Delta \zeta = - \frac{\tau_b \Delta l}{\rho_w g h}, \quad (3)$$

since the first term of the right side in Eq. 1 equals the fourth term of the same side. As to the bottom friction the quadratic formula $\tau_b = \rho_w \gamma_w^2 u^2$ is usually used, where a dimensionless constant γ_w^2 is of the order of 2×10^{-3} .

First consider the outflow of river water as the constant current. The volume transport from rivers to the bay is not usually abundant in our country, and a mean flow exceeding 1 cm/s can not be expected when the river water is assumed to run uniformly through the mouth of the bay. If we put $\Delta l = 50$ km, $h = 20$ m and $u = 1$ cm/s we have $\Delta \zeta = 0.005$ cm only from Eq. 3. The difference in sea level is also small in the case when the river water is limited within a surface layer as generally observed. On the other hand, in some of the bays, a weak circulation is found. Even if we assume its velocity 10 cm/s throughout

the greater part of the bay, which seems to be overestimated, we have only 0.5 cm as the difference in sea level in the basin of the above assumed dimension. By the way, when a tidal station stands by a river of fast flow, rather steep slope of water surface will be observed. For instance, at a station 500 m up from the mouth of the river where the depth is 5 m and the stream 1 m/s the mean level at the station is higher 2 cm than that at the mouth of the river.

B. *Effect of meteorological elements*

The sea surface of a small basin will incline in the same direction as the mean wind over years. In this case the first, the third and the last terms in the right side of Eq. 1 become important and we get

$$\Delta\zeta = \frac{(\tau_s - \tau_b)\Delta l}{\rho_w g h} \quad (4)$$

Tangential stress on the surface is usually given by $\tau_s = \rho_a \gamma_a^2 W^2$, where ρ_a is the density of air, W the speed of wind and γ_a^2 nearly a constant of the order of 2×10^{-3} . The bottom stress may be ten percent or so of the surface stress.

In our country the northerly wind prevails in general as understood from the fact that mean direction of wind calculated by frequency is N9°W in Tokyo and N2°W in Osaka. Therefore the mean wind is likely to cause descent of sea surface near the inner end of the basin. The mean of absolute values of wind velocity, usually called the mean velocity of wind, is several meters per second on the surface of the bay, but the mean of wind velocity as a vector is considerably smaller than this. If we assume the mean wind of 3 m/s, the difference in sea level at a distance of 50 km is less than 1 cm in a basin with the depth 20 m. In actual bays with an open end the difference of sea level will be much smaller than that in closed basins.

Next, when the mean atmospheric pressure is not uniform over the bay, the difference of sea level is also found. From the first and the second terms in the right side of Eq. 1

$$\Delta\zeta = -\frac{\Delta p_s}{\rho_w g} \quad (5)$$

is understood for the difference of atmospheric pressure Δp_s . For example, Tokyo is in the mean atmospheric pressure 0.7 mb higher than Mera at the top of Boso Peninsula, so the mean sea level is 0.7 cm lower at the bay bottom than at the mouth.

C. Effect of Coriolis force

On the rotating earth the sea surface slopes perpendicularly to the axis of flow by the action of Coriolis force. Taking x in the direction of the current and y in the normal to it, as the difference in sea level between two points at intervals of Δs across the current we have

$$\Delta\zeta = -\frac{fu\Delta s}{g} \quad (6)$$

from a balance of the first term of the right side in Eq. 2 with the last term of the left side. This effect is small in a small basin where a constant current is weak and does not cover a wide area. For example, in order that the difference of sea level of 1 cm is produced at a breadth of 20 km, it is required that a rather conspicuous constant current with the speed of about 7 cm/s flows uniformly through the breadth of 20 km ($f=0.73\times 10^{-4}$ sec $^{-1}$ in the vicinity of Japan), but such a case will not be recognized in our bays in general. On the open sea, however, remarkable difference in sea level is expected in the region of a strong ocean current. If we put $u=1.2$ knots and $\Delta s=60$ sea miles we have $\Delta\zeta=50$ cm across the current (cf. Fig. 3).

D. Effect of centrifugal force

In a region where a current curves, the sea surface inclines radially by the centrifugal action. Let the radius of curvature be r and the speed of the current q . The difference of sea levels on both sides of a bay with the breadth of Δr is

$$\Delta\zeta = \frac{q^2\Delta r}{gr} \quad (7)$$

Since it depends on the square of the speed, inclination of mean surface arises not only in case of a constant current but also in case of an oscillatory current such as the tidal current. Such circumstances will be expected in both Bays of Ariake and Ise from their shape (Fig. 5). In Ariake Bay $\Delta\zeta$ becomes 1.2 cm when we put $r=20$ km, $\Delta r=15$ km and $u=40$ cm/s as a root mean square of the tidal current. In Ise Bay $\Delta\zeta$ becomes 0.4 cm when we put $r=25$ km, $\Delta r=25$ km and $u=20$ cm/s in the same sense as above. This effect is recognized in Figs. 10 and 11 showing the mean sea level in both bays. The figures, however, are based on numerical computations.

E. Effect of ununiform distribution of density of sea water

Relative height of sea surface will be recognized in a region where the density of sea water is not uniformly distributed. If we assume that the density is constant with depth and the hydrostatic pressure is horizontally balanced at the depth of d , the difference in sea level between two points becomes

$$\Delta\zeta = -\frac{d\Delta\rho_w}{\rho_w} \quad (8)$$

for the difference of density of $\Delta\rho_w$. An example, showing the annual variation of density on the surface layer at three places, is indicated in Fig. 6. (The

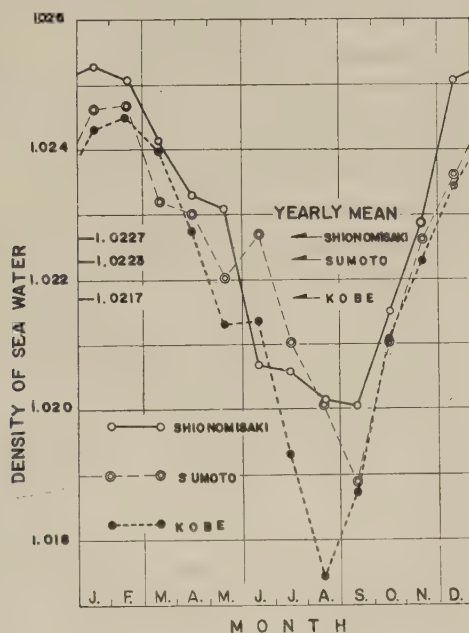


Fig. 6. Annual variation of the density of sea water at three stations (unit, gr/cm³). Means were taken over 1941-48 at Shionomisaki, 1948-50 at Sumoto and 1946-48 at Suma, Kobe.

density was calculated using observed results of water temperature and σ_{15}). Of the three stations, Kobe and Sumoto are located near the inner end and the open end of Osaka Bay respectively, and Shionomisaki close to Kushimoto is at the top of Kii Peninsula and faces to the open sea directly.

The biggest difference in monthly mean density is found in August with the value of 0.0027 between Kobe and Shionomisaki. The yearly mean of density is

1.0217 at Kobe, 1.0223 at Sumoto and 1.0227 at Shionomisaki, and the difference is only 0.001 between Kobe and Shionomisaki. Accordingly it will be sufficient if we take $\Delta\rho_w=0.001$ as the difference in density between the inner end and the entrance of usual bays. Assuming $d=10$ m, though its decision is difficult and arbitrary somewhat, we get $\Delta\zeta=1$ cm. In case of $d=20$ -m we get $\Delta\zeta=2$ cm and so on. This suggests that ununiform distribution of density will be considerably effective in rather deep bays, but in usual shallow bays we can hardly expect the result indicated in Fig. 5 from it alone. Further discussions, however, are difficult on account of lack of materials.

Moreover, the effect of wave set-up (LONGUET-HIGGINS and STEWART, 1964) may be suggested. But we can easily understand that it will be out of the question, considering the observed fact that waves higher than 2 m occur with frequency of only two percent in a year at Tagonoura situated in the inner end of Suruga Bay which seems to be the roughest among several bays in question.

So far we have discussed the effect of various factors which might influence mean sea level in a bay, but definitely effective factors which can interpret the result shown in Fig. 5 were not found out among them. Hence we intend to introduce a new factor in the following sections.

4. Mean sea level of the second order due to the standing oscillation of tide

That the mean sea level inclines accompanied with the standing oscillation of tide was stated by the present authors at the meeting of Oceanographical Society of Japan in 1964, in order to explain the remarkable rise of sea surface in Ariake Bay presented in Fig. 4 (ISOZAKI and UNOKI, 1964). The method will be shown below. On the other hand, in case of free standing oscillation, the similar result was also gained by LONGUET-HIGGINS and STEWART in the same year, applying the idea of radiation stress (LONGUET-HIGGINS and STEWART, 1964).

Consider a long wave of one dimension for simplicity. When we neglect v , p_s and τ_s in Eq. 1, we have

$$\frac{\partial u}{\partial t} + u \frac{\partial u}{\partial x} = -g \frac{\partial \zeta}{\partial x} - \frac{\gamma_w^2}{D} u |u| \quad (9)$$

as the equation of motion. The equation of continuity is

$$\frac{\partial \zeta}{\partial t} = -\frac{\partial}{\partial x} (Du). \quad (10)$$

In a partly closed waterway, standing waves are produced and we may assume such a solution as

$$\left. \begin{aligned} \zeta(x, t) &= \bar{\zeta}(x) + \zeta_I(x, t) + \zeta_{II}(x, t) + \cdots \\ u(x, t) &= u_I(x, t) + u_{II}(x, t) + \cdots, \end{aligned} \right\} \quad (11)$$

where $\bar{\zeta}$ is the mean of elevation with time, ζ_I and u_I present the elevation and flow of standing waves which satisfy the linearized equations, ζ_{II} and u_{II} higher harmonics of the second order and so on. Mean velocity $\bar{u}$ disappears on the present condition that the state of the wave remains constant with time. Mean with time is indicated by a bar above variables. Substituting Eq. 11 into Eq. 9 and neglecting small terms of higher order than the second, we get

$$\frac{\partial}{\partial x} \left(\frac{1}{2} \overline{u_1^2} \right) = -g \frac{\partial \bar{\zeta}}{\partial x}$$

after average with time was taken. By integrating this we have

$$\bar{\zeta} + \frac{1}{2g} \overline{u_1^2} = \text{const.} \quad (12)$$

On the other hand, $\overline{\zeta_I u_I} = 0$ is demanded by substituting Eq. 11 into Eq. 10 and averaging it. It is naturally allowable in case of standing oscillations in question. We can easily understand from Eq. 12 that the mean water surface is high in regions where the current is weak and low in regions where it is strong.

As the first example consider a rectangular closed basin of uniform depth h and length l indicated in Fig. 7-I. The solution of the first order, which satisfies

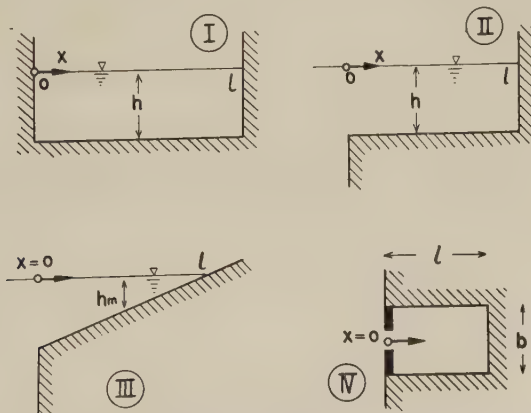


Fig. 7. Four model basins, of which I, II and III indicate cross sections and IV a plane figure.

the linearized equation $\left(\frac{\partial u}{\partial t} = -g \frac{\partial \zeta}{\partial u} \text{ and } \frac{\partial \zeta}{\partial t} = -h \frac{\partial u}{\partial x}\right)$ and the boundary conditions $((u)_{x=0} = (u)_{x=l} = 0)$, is

$$\zeta_1 = 2\alpha \cos kx \cos(\sigma t + \varepsilon) \quad (13)$$

$$u_1 = \frac{2\alpha C}{h} \sin kx \sin(\sigma t + \varepsilon) \quad (14)$$

as well known, where 2α is the amplitude, σ the angular velocity and C the speed of long wave $\sqrt{gh}$. The wave number k is given by $k = n\pi/l$, $n=1, 2, \dots$, and such a relation as

$$k = \frac{\sigma}{C} \quad (15)$$

holds in general. From Eq. 14 we have $\overline{u_1^2} = (2\alpha^2 C^2 / h^2) \sin^2 kx$. Putting it into Eq. 12 and considering the condition that $\int_0^l \bar{\zeta} dx = 0$, the mean surface level becomes

$$\bar{\zeta} = \frac{\alpha^2}{2h} \cos 2kx, \quad (16)$$

which just accords with the result of LONGUET-HIGGINS and STEWART (1964). When the depth of the basin is not sufficiently small compared with the wave length, LONGUET-HIGGINS and STEWART gave a solution such as

$$\bar{\zeta} = \alpha^2 k \coth 2kh \cos 2kx. \quad (17)$$

It is seen that this result is gained only if we multiply the result of long wave, Eq. 16, by a constant $2kh \coth 2kh$. Therefore we will consider the case of long wave only hereafter.

4.1 The case of a bay with uniform depth

We will investigate the mean sea level accompanied with the cooscillating tide in a bay of length l and uniform depth h (Fig. 7-II). On the condition at the mouth such as

$$\zeta(o) = a \cos(\sigma t + \varepsilon), \quad (18)$$

the cooscillating tide of the first order is given by

$$\zeta_1 = a \frac{\cos k(l-x)}{\cos kl} \cos(\sigma t + \varepsilon) \quad (19)$$

$$u_1 = -\frac{aC}{h} \frac{\sin k(l-x)}{\cos kl} \sin(\sigma t + \varepsilon), \quad (20)$$

where the wave number k is determined from given values of σ and h by Eq. 15. Substituting Eq. 20 into Eq. 12 and assuming $\zeta(0)=0$ at the bay mouth, we have

$$\zeta(x) = \frac{a^2}{8h \cos^2 kl} \{ \cos 2k(l-x) - \cos 2kl \} . \quad (21)$$

At the bay bottom it becomes

$$\zeta(l) = \frac{a^2}{4h} \tan^2 kl . \quad (22)$$

The right of Fig. 8 presents the mean sea level at the bay bottom for various values of kl and the left figure the shape of mean sea level in the basin for special cases where kl is 0.5 and 1.5.

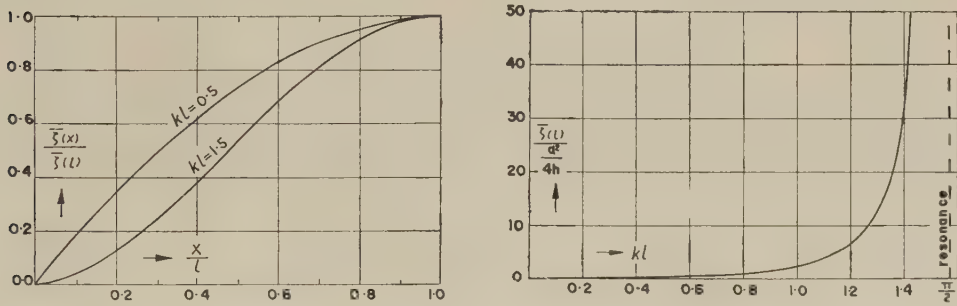


Fig. 8. Ascent of mean sea level in a partly closed basin of uniform depth (Fig. 7-II).
Right: Mean sea level at the bay bottom, based on Eq. 22. Left: Configuration of sea surface, based on Eq. 21.

The sea level becomes high towards the bay bottom. In particular case of $kl = \frac{\pi}{2}$, which corresponds to the case that the period of external wave ($2\pi/\sigma$) is equal to that of fundamental proper oscillation in the bay ($4l/C$), a remarkable standing oscillation is generated on account of resonance, and consequently the noticeable rise of mean sea level is also raised. If we put $l=90$ km, $h=20$ m, $a=1.7$ m and $2\pi/\sigma=12$ hours, roughly corresponding to the spring tide in Ariake Bay, we have $\zeta(l)=6.4$ cm from Eq. 22. It is considerably smaller than the value recognized in Fig. 4. This will be attributed to the fact that the depth and configuration of the basin is assumed to be too simplified. The result of numerical computations where the actual geographical features are taken into consideration will be shown later in Section 4.3.

4.2 The case of a bay with constant slope of the sea-bottom

The mean sea level in the inside of the bay is not always higher than that at the mouth as understood from Eq. 12. As an example we will take a bay with constant slope of bottom (Fig. 7-III), where the depth is presented by $h=2h_m(1-x/l)$ when the mean depth is h_m . We consider Eq. 18 as the boundary condition at the mouth similarly to the previous case. The standing oscillation in linear case is given by

$$\zeta_I = a \frac{J_0\{2^{1/2}k_*l(1-x/l)^{1/2}\}}{J_0(2^{1/2}k_*l)} \cos(\sigma t + \varepsilon) \quad (23)$$

$$u_I = -\frac{aC_m}{2^{1/2}h_m(1-x/l)^{1/2}} \cdot \frac{J_1\{2^{1/2}k_*l(1-x/l)^{1/2}\}}{J_0(2^{1/2}k_*l)} \sin(\sigma t + \varepsilon), \quad (24)$$

where $J_0(\xi)$ and $J_1(\xi)$ mean the Bessel functions of the zero order and the first order respectively, and

$$C_m = \sqrt{gh_m}, \quad k_* = \sigma / \sqrt{gh_m} \quad (25)$$

(for example, LAMB, 1932). The solution satisfies the condition that the water oscillates along the slope of the coast at the inmost end of the basin. Substituting Eq. 24 into Eq. 12 and assuming $\zeta(0)=0$, the mean sea level becomes

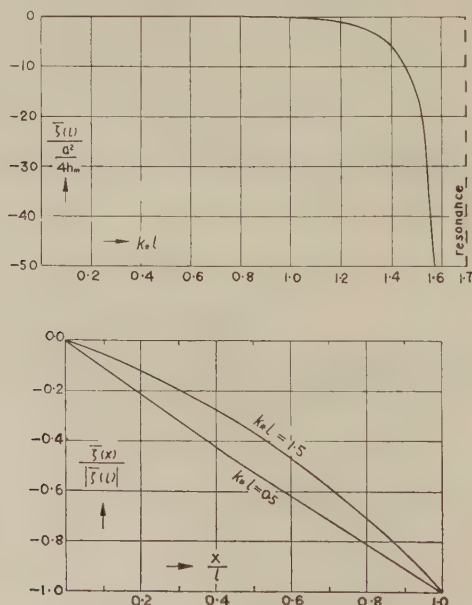


Fig. 9. Descent of mean sea level in a partly closed basin of constant bottom slope (Fig. 7-III). Upper: Mean sea level at the bay bottom, based on Eq. 27. Lower: Configuration of sea surface, based on Eq. 26.

$$\bar{\xi}(x) = \frac{a^2}{8h_m} \frac{\{J_1(2^{1/2}k_*L)\}^2 - \{J_1(2^{1/2}k_*L(1-x/L)^{1/2})\}^2 / (1-x/L)}{\{J_0(2^{1/2}k_*L)\}^2}. \quad (26)$$

At the bay bottom we have

$$\bar{\xi}(l) = \frac{a^2}{8h_m} \cdot \frac{\{J_1(2^{1/2}k_*L)\}^2 - \frac{1}{2}(k_*L)^2}{\{J_0(2^{1/2}k_*L)\}^2}. \quad (27)$$

The height of mean sea level at the bay bottom and the configuration of the sea surface are indicated in the upper and the lower of Fig. 9 respectively. The mean sea level declines towards the bay bottom in contrast with the case of uniform depth and the lowering becomes conspicuous when the condition of resonance is satisfied ($k_*L=1.70$). Therefore it should be kept in mind that the distribution of mean sea level differs remarkably according to the geographical features of the bay in question.

4.3 The result of numerical computations

A reasonable agreement between the theoretical value and observed one of mean sea level was not gained by assuming a simple model basin of one dimension as stated above. So it becomes necessary to take the actual geographical features of the basin into consideration as much as possible. To do it we compute the tide numerically first and calculate the mean sea level next. As the equations of motion for the tide, Eqs. 1 and 2 are used after removing p_s and τ_s . In order to solve these equations as well as the equation of continuity numerically, they were transformed into finite difference equations with a time difference Δt and a space difference Δs . And then, the flow and elevation of tide were computed at respective grid points step by step, using the difference equations of motion and continuity alternatively. Details of the procedure of computation are stated in another report of the authors (UNOKI, ISOZAKI and OTSUKA, 1964). Computation were made for three bays such as Tokyo Bay, Ise Bay and Ariake Bay. We used $2\Delta t=60$ sec and $2\Delta s=3.0$ km for Tokyo Bay, $2\Delta t=60$ sec and $2\Delta s=4.0$ km for Ise Bay, and $2\Delta t=90$ sec and $2\Delta s=3.5$ km for Ariake Bay. The mean sea level at some selected grid points was obtained by averaging the calculated elevations over one period of tidal oscillation.

In case of Ariake Bay we took a semi-diurnal tide assuming the amplitude of 145 cm at the entrance, which nearly corresponds to the spring tide. The distribution of computed mean sea level is presented in the left of Fig. 10. The abrupt rising of mean sea level from the mouth towards the central region

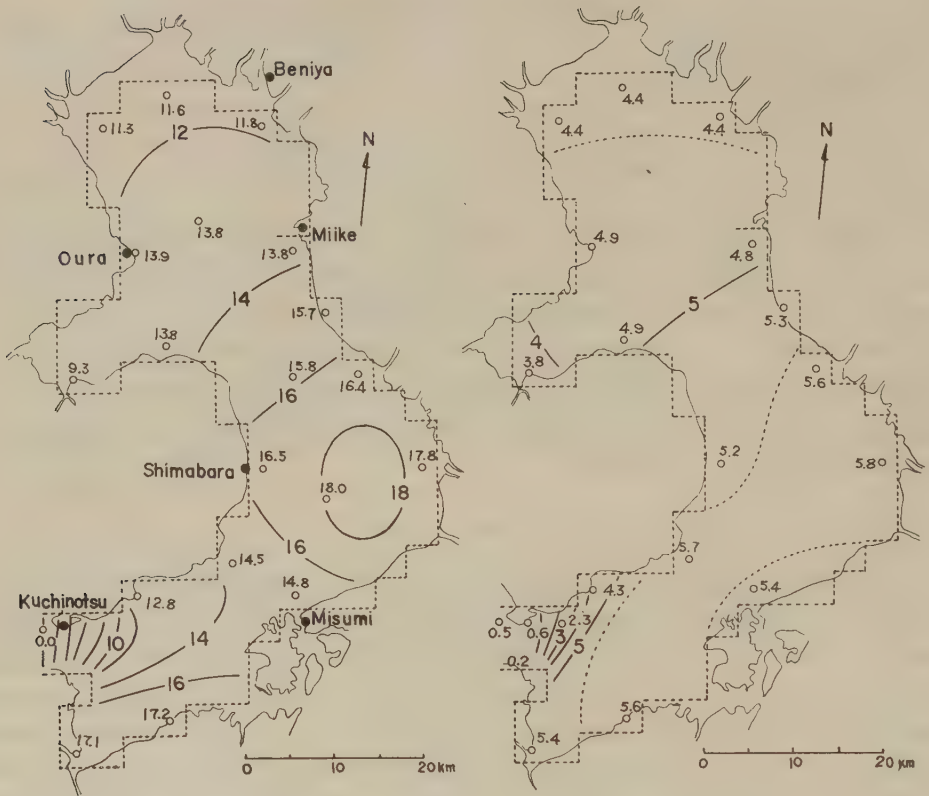


Fig. 10. Theoretical distribution of mean sea level in Ariake Bay (unit, cm). Numerical computations were done on the condition that the amplitude of tide at the mouth of bay is 145 cm for the left case and 70 cm for the right case, but the period is 12.421 hours in each case.

accords with what is shown in Fig. 4, but the amount of rising is 18 cm which corresponds to sixty percent of the observed value. Towards the bay bottom from the central region, the sea level descends gradually on the contrary from the observed result indicating gentle ascent. However, considering various errors contained in the observed values and conditions of the computation, we may roughly say that the majority of remarkable rising of mean sea level is due to the inclination of sea surface caused by the predominant standing oscillation of tide. We can also recognize in the figure that the level is higher on the east coast than on the west coast on the whole, and it is especially conspicuous in the region where the current is strong and its path makes a sharp curve. This

is of course attributed to the effect of centrifugal force as discussed already in Section 3D.

A case when the amplitude at the bay mouth becomes about a half ($a=70$ cm) was also computed with the result shown in the right of Fig. 10. The magnitude of mean sea level is about one third of the value computed first. It is a little larger than one fourth which is expected, for instance, from Eq. 21 including square of the amplitude. This small discrepancy may partly show the effect of other terms of higher order than the second, which was not taken in introducing Eq. 21.

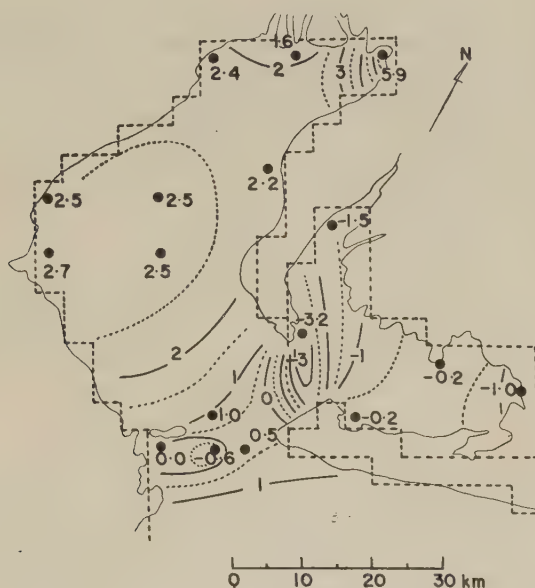


Fig. 11. Theoretical distribution of mean sea level due to the tide in Ise Bay on the condition that the amplitude at the mouth is 90 cm and the period 12 hours (unit, cm).

Fig. 11 represents the theoretical distribution of mean sea level in Ise Bay nearly at the time of spring tide. The level at Nagoya, innermost of the Bay, is about 6 cm higher than the level at Toba located at the mouth. It corresponds to a half of the value indicated in Fig. 5. Pretty errors, however, may be possibly contained in the latter value, because the settlement of ground is violent in the region around the bay bottom (Fig. 2). We can faintly recognize the effect of centrifugal force in this bay, too. It is interesting that in a small bay surrounded by two prolonged peninsulas (Mikawa and Atsumi Bay) the sea

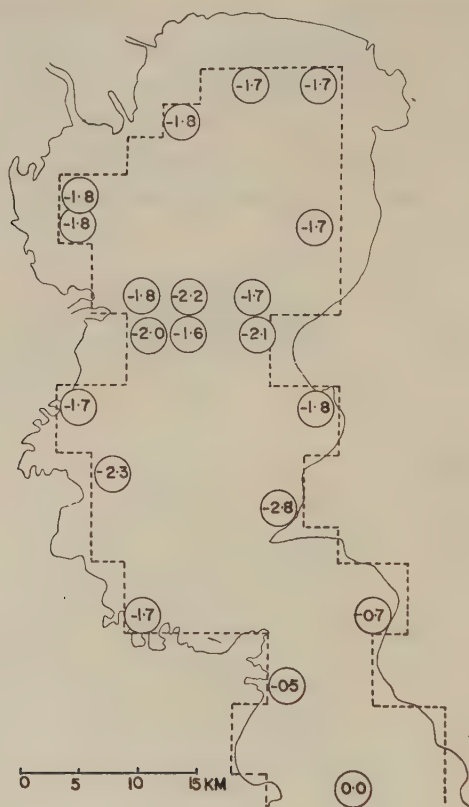


Fig. 12. Theoretical distribution of mean sea level due to the spring tide in Tokyo Bay (unit, cm).

As for Tokyo Bay, according to Fig. 12 showing the result of numerical computations at the time of spring tide, the mean sea level in the inside is lower than that at the entrance though the difference is very small. On the contrary, however, the published data suggested that the level is higher a little inside than outside with a small difference. But we feel difficult to arrive at definite conclusion unless more reliable data are gained, because these differences are near the possible errors contained in the actual data.

4.4 Discussions

From the result of numerical computations mentioned above, we may say

that the greater part of ascent of mean sea level found in both Bays of Ariake and Ise is due to the standing oscillation of tide, while in case of Tokyo Bay we could not come to definite conclusion when the errors of observed data were taken into consideration.

In the Bays of Suruga and Kagoshima, the slope of mean sea level due to the tide seems to be negligibly small because of deep basin. Namely, when we put $h=50$ m, $l=60$ km and $a=105$ cm for Kagoshima Bay, Eq. 22 gives $\xi(l) \doteq 0.1$ cm, and in case of Suruga Bay where the mean depth is several hundred meters it is further smaller.

The rise of sea level due to the tide can not be expected in Nagasaki Harbor, either, because of small dimension of the basin. On the other hand, Nagasaki Harbor is famous for the seiches called *abiki* there develop and gave damages sometimes (UNOKI, 1959). Now consider a basin shown in Fig. 7-II. In the basin, such a seich as

$$\xi_1 = 2\alpha \sin kx \sin(\sigma t + \varepsilon) \quad (28)$$

$$u_1 = \frac{2\alpha C}{h} \cos kx \cos(\sigma t + \varepsilon) \quad (29)$$

is generated. Here C is $\sqrt{gh}$ and $k=(2n-1)\pi/2l$, $n=1, 2, \dots$. Substituting Eq. 29 into Eq. 12 and assuming $\xi(0)=0$, the mean sea level accompanied with the seich becomes

$$\bar{\xi}(x) = \frac{\alpha^2}{2h} (1 - \cos 2kx), \quad (30)$$

and at the closed end we have

$$\bar{\xi}(l) = \frac{\alpha^2}{h}. \quad (31)$$

Therefore, for instance, for $\alpha=50$ cm and $h=8$ m we have $\bar{\xi}(l) \doteq 3$ cm. But this effect on yearly mean sea level is very faint because big seiches do not occur so frequently even in Nagasaki Harbor, as known from the observed data that seiches higher than 50 cm were generated only seventy times or so for twenty one years (UNOKI, 1959).

After all, in bays other than Ariake and Ise, we could not sufficiently interpret the rising of mean sea level suggested in Fig. 5, though various factors were taken into consideration. However, some observed data seem to contain considerable errors and this makes it difficult to discuss in detail. If in the region around a bay bottom the settlement of the ground, as seen in Fig. 2, happens between the period for levelling and that for mean sea level, it will be apparently observed that the mean sea level is higher at the inner end than at

the entrance. Accordingly it remains a subject in future to investigate whether the results in these bays are really true or not and, if they are true, why they are. As to the ascent of mean sea level in the Seto Inland Sea (Fig. 3) and Osaka Bay, the authors think that it must be attributed to the slope of sea level accompanied with the tide, though numerical computations are not done yet.

On the other hand, we can also expect temporary rising (or falling in some case) of mean sea level in bays where a great tsunami is sweeping, because the condition of resonance in question is likely to be satisfied.

4.5 A supplementary note on the effect of a dike on mean sea level

Recently in our country, it is proposed to construct a great breakwater with a constricted opening across important bays for the purpose to reduce storm surges as well as sea waves. In such a case the state of tidal oscillation is likely to vary more or less, and consequently the distribution of mean sea level will also vary, especially greatly near the narrow opening of the breakwater where non-linearity is predominant.

Let's consider a simple basin (Fig. 7-IV) of length l , breadth b and uniform depth h , being connected with an outer sea through a narrow opening of breadth b_* and depth h_* . The velocity of current is written by u and u_* for the inside and the opening respectively. We consider the elevation at just outside of the opening to be

$$\zeta(-0) = a \cos \sigma t, \quad (32)$$

and assume that the current at the entrance is given by

$$u_* = u(+0) \frac{bh}{b_*h_*} = \pm \kappa \sqrt{2g|\zeta(-0) - \zeta(+0)|}, \quad (33)$$

where κ is a discharge coefficient, $\zeta(+0)$ and $u(+0)$ the elevation and the velocity in just the rear of the opening and the signs $\pm$ correspond to $\zeta(-0) \gtrless \zeta(+0)$ respectively. According to the result gained by LOVE (1959), the solution of the first approximation is

$$\zeta_1(x, t) = a \cos \theta \frac{\cos kl(1-x/l)}{\cos kl} \cos(\sigma t + \theta) \quad (34)$$

$$u_1(x, t) = -\frac{aC}{h} \cos \theta \frac{\sin kl(1-x/l)}{\cos kl} \sin(\sigma t + \theta), \quad (35)$$

where

$$k = \sigma / \sqrt{gh}, \quad \sin \theta = \{1 \pm \sqrt{1 + 4B^2}\} / 2B \quad \text{and} \quad B = \frac{4}{3\pi} \frac{a\sigma^2}{g} \left(\frac{bl}{\kappa b_* h_*} \right)^2 \left(\frac{\tan kl}{kl} \right)^2.$$

Substituting Eq. 35 into Eq. 12, the mean sea level inside becomes

$$\bar{\zeta}(x) = \frac{a^2}{4h} \tan^2 kl \cos^2 \theta \left[\left(\frac{bh}{b_* h_*} \right)^2 - \left(\frac{\sin kl(1-x/l)}{\sin kl} \right)^2 \right] \quad (36)$$

above the mean sea level at the outside of the opening.

At the inner end we have

$$\bar{\zeta}(l) = \frac{a^2}{4h} \tan^2 kl \cos^2 \theta \left(\frac{bh}{b_* h_*} \right)^2, \quad (37)$$

which differs from Eq. 22 in an additional coefficient $\cos^2 \theta (bh/b_* h_*)^2$. It is noticeable that a gap of mean sea level such as

$$\bar{\zeta}(+0) = \frac{a^2}{4h} \tan^2 kl \cos^2 \theta \left[\left(\frac{bh}{b_* h_*} \right)^2 - 1 \right] \quad (38)$$

is found between both sides of the constricted opening.

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Water Characteristics of the Kuroshio*

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Abstract: A typical hydrographic section of the Kuroshio south of Honshu is indicated from the observation made in May 1965 on board the *Ryofu Maru*. The geostrophic fluxes of the Kuroshio in each class with temperature range of 1°C and with salinity range of 0.2‰ have been computed from the typical section and six sections south of Cape Shiono-misaki, three in summer and three in winter. The differences of the Kuroshio water characteristics between winter and summer are described and the water characteristics of the Kuroshio south of Japan are discussed in comparison with the North Equatorial Current and with the Kuroshio Extension east of Japan.

1. Introduction

A hydrographic section south of Honshu has been often chosen as a representative section of the Kuroshio, and a large number of hydrographic sections have been made by many agencies and institutes. Unfortunately, however, no comprehensive descriptions of water characteristics carried by the Kuroshio have been made, despite lots of descriptive papers of the Kuroshio south of Japan.

The present short note has been made to get an idea of water characteristics of the Kuroshio. A deep hydrographic section has been used to present a typical distribution of water properties across the Kuroshio. Six hydrographic sections which are relatively long, deep and reliable in accuracy enough to be described have been chosen south of Cape Shiono-misaki to see differences between winter and summer. The gross features of water characteristics of the North Equatorial Current, the Kuroshio, and the Kuroshio Extension are demonstrated.

2. A typical hydrographic section of the Kuroshio south of Honshu

A deep hydrographic section of the Kuroshio was made along 136.5°E long. in May 1965 on board the *Ryofu Maru*, Japan Meteorological Agency. The

* Currents and water characteristics of the Kuroshio System (XIII)

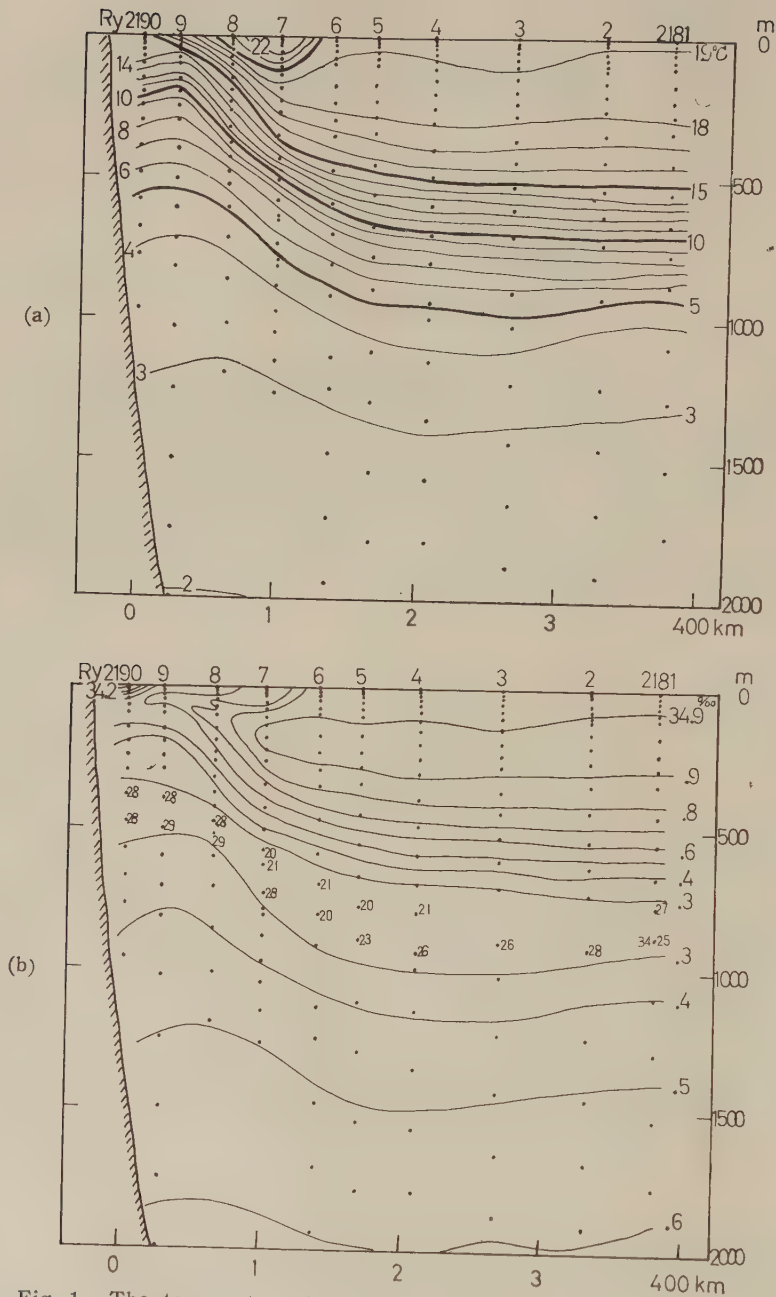
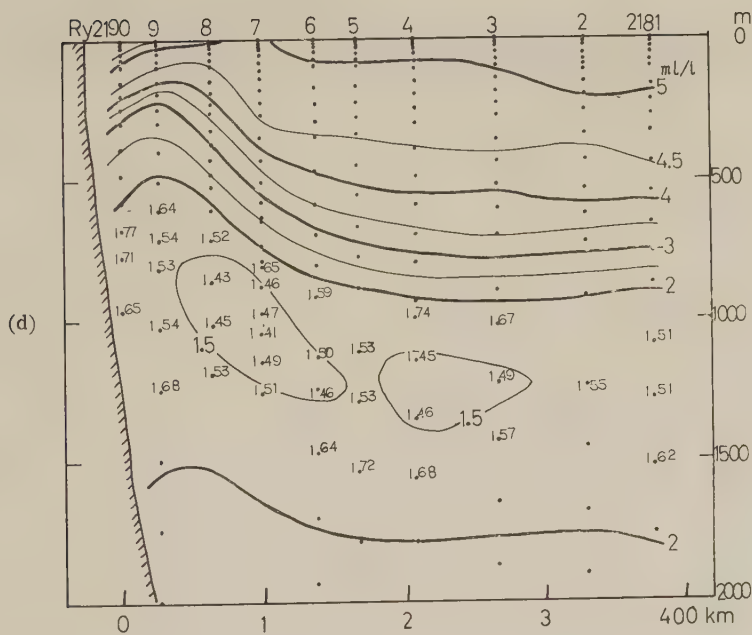
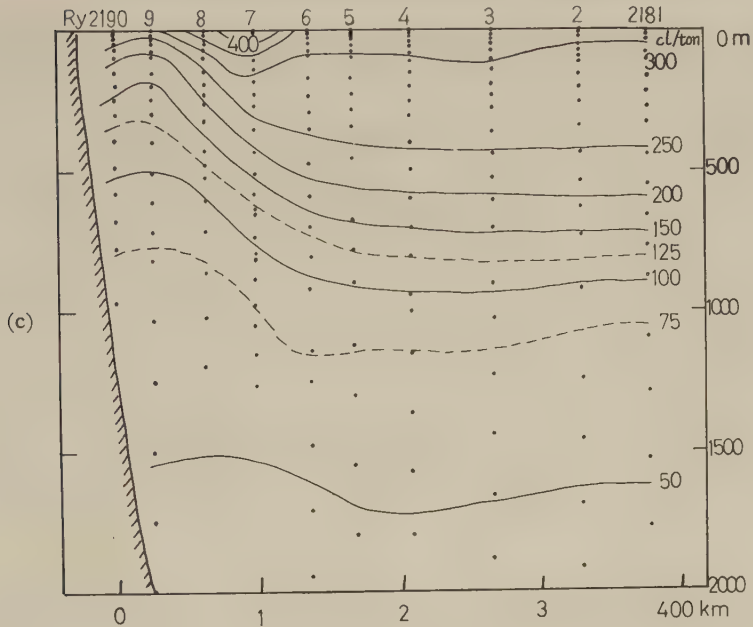


Fig. 1. The temperature (a), salinity (b), thermobaric anomaly (c), and oxygen (d) section across the Kuroshio at 136.5°E long., May 1965, Ryofu Maru (vertical exaggeration, 200).

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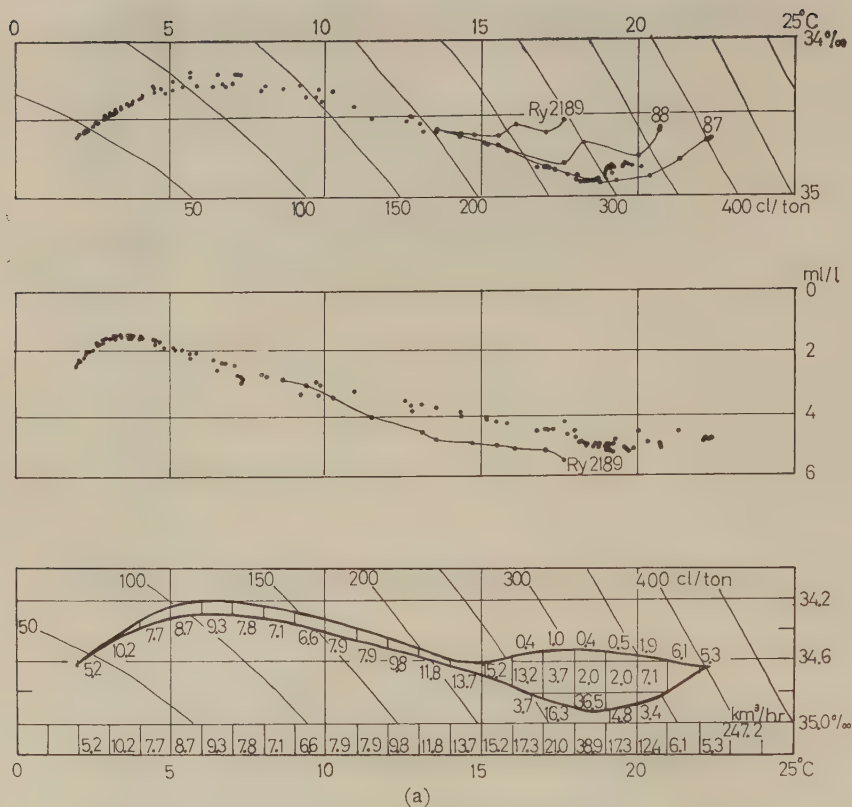
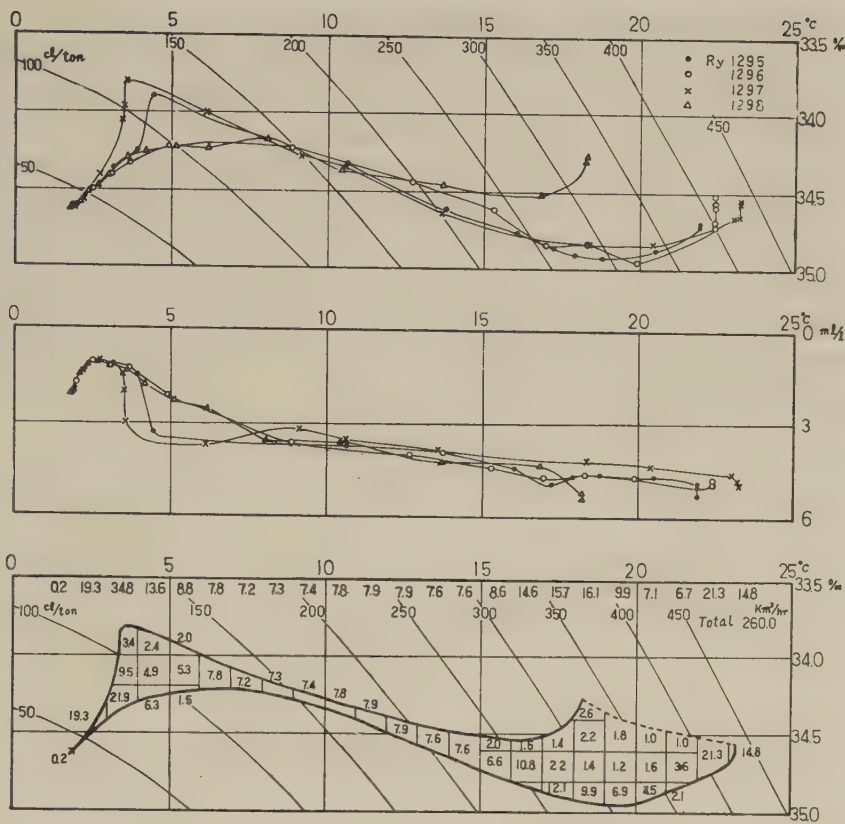


Fig. 2. Temperature-salinity relations, temperature-oxygen relations, and geostrophic fluxes (km^3/hr) on T-S diagram of the Kuroshio, south of Japan, 136.5°E , May 1965 (a), east of Japan, 144°E , November 1959 (MASUZAWA, 1964 b) (b).

section was nearly perpendicular to the general direction of the Kuroshio current. It may give a satisfactory information of typical cross-current structure of the Kuroshio.

The temperature section (fig. 1a) indicates such familiar features in the high-velocity zone, as 'warm core', the warmest water in the surface layer and the inclined main-thermocline ($6\text{--}16^\circ\text{C}$) at the mid depth. The salinity-minimum layer centered at the surface of thermosteric anomaly 125 cl/ton can be found in the lower part of, or just below the main thermocline. The minimum salinity seems to be rather larger at the left side of the Kuroshio current-axis than at the high-velocity zone (fig. 1b). The oxygen-deficient layer, say less than 2 ml/l , is as thick as 800 m . The core is on the surface of thermosteric anomaly

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(b)

70 cl/ton or so, and the minimum value seems to be somewhat less than 1.5 ml/l in the current-zone of the Kuroshio.

3. Water characteristics of the Kuroshio south and east of Japan

The geostrophic flux of the Kuroshio has been computed between St. 2184 and St. 2189 in the 136.5°E-section mentioned, assuming to be motionless at 2000 db. The total flux is as much as 247 km³/hr (69×10^6 m³/s). The procedure for computation of the distribution of geostrophic flux on the T-S diagram is the same as used in the previous paper (MASUZAWA, 1964b). The geostrophic flux through two adjacent stations of the section is calculated every 50 m thickness in the upper 1000 m layer, and every 100 m in the lower 1000 m layer. The computed flux is divided into each class of water characteristics with temperature range of 1°C and with salinity range of 0.2‰ and the area of each class is

estimated from isotherms and isosalines on the section.

The most outstanding mode in the distribution of flux on the temperature-salinity diagram can be found in the layer of 17-19°C, centered at the class of 18.5°C, 34.85‰ (fig. 2a). The water is equivalent to the so-called '18° water' which is seen extensively in the western subtropical region of the North Pacific and of the North Atlantic (WORTHINGTON, 1959). The temperature-salinity relation below the surface of thermoclinic anomaly 250 cI/ton is quite uniform over the section, and the distribution of flux against temperature below the surface is also fairly uniform. It may be noticeable that the flux of 18° water decreases and the flux of less-saline, oxygen-rich intermediate water increases in the Kuroshio Extension east of Japan (fig. 3).

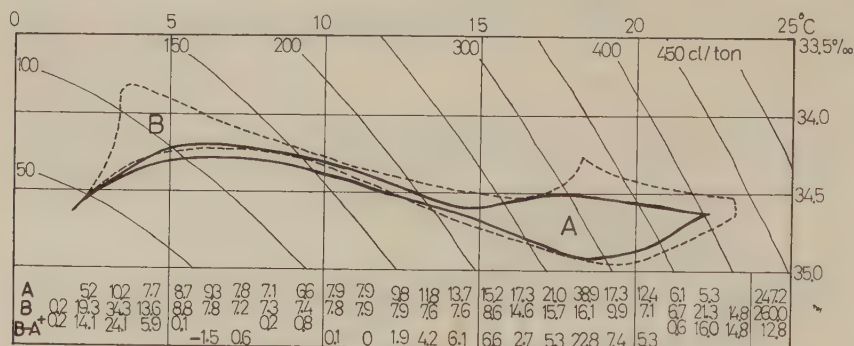


Fig. 3. Difference of T-S relation and geostrophic flux (km³/hr) of the Kuroshio between south of Japan (A) and east of Japan (B).

4. The Kuroshio water characteristics south of Cape Shiono-misaki in summer and in winter

Six good hydrographic sections made south of Cape Shiono-misaki have been chosen to see some differences of water characteristics of the Kuroshio between winter and summer (table 1).

The method of computation in this case is nearly the same as used by MASUZAWA (1964a) in the description of the Pacific North Equatorial Current. The north, or south limit of the Kuroshio is assumed to lie at the station where the geopotential at the sea surface is minimum, or maximum, respectively. The hydrostatic computation has been done from the reference level of 1000 db. The geostrophic flux is divided into each characteristic class with temperature range of 1°C and with salinity range of 0.2‰. The water characteristics at a station

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Table 1. Geostrophic computation in six sections of the Kuroshio south of Cape Shiono-misaki.

Date	Vessel	$\Delta\phi^{(a)}$ (dyn. m)	Flux (km ³ /hr)	T ^(b) (°C)	S ^(b) (‰)	Flux (km ³ /hr)	Salt flux (ton/s)
4-5, Sep., 1951	Chikubu Maru	0.81	158	17.05	34.48	178	1.75×10^6
9-10, Aug., 1953	Ryofu Maru	0.93	175				
28-29, Aug., 1955	Shumpu Maru	0.99	201				
10-11, Mar., 1953	Shiga Maru	1.07	209	15.45	34.63	197	1.96×10^6
19-20, Mar., 1955	Shumpu Maru	0.93 ^(c)	182 ^(c)				
18-19, Feb., 1956	Meiyo Maru	1.03	201				

- (a) Geopotential difference at the sea surface between both limits of the Kuroshio; corrected by the empirical relationship between vertically integrated thermosteric anomaly and geopotential anomaly (YOSHIDA & TSUCHIYA, 1957).
(b) Average weighted by flux.
(c) The section seems too short to cover the south limit of the Kuroshio.

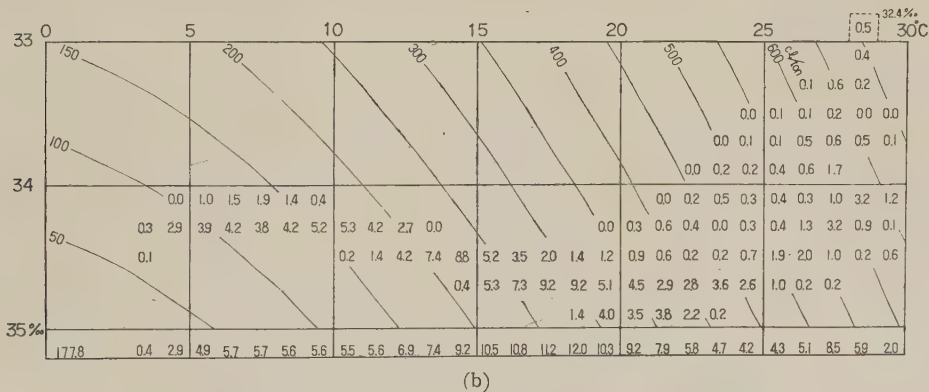
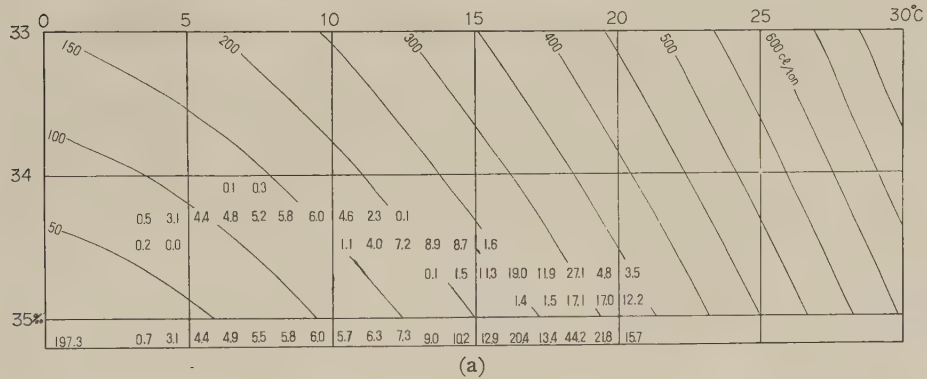


Fig. 4. The average geostrophic fluxes (km³/hr) of three winter sections (a) and of three summer sections (b) south of Cape Shiono-misaki.

are assumed to have half the geostrophic flux computed from two pairs of adjacent stations including the station. The vertically integrated thermosteric-anomaly is assumed as an approximation of geopotential anomaly (MONTGOMERY & WOOSTER, 1954; YOSHIDA & TSUCHIYA, 1957).

The distribution of flux on a temperature-salinity diagram has been made for each of the six sections, but the results do not seem reliable enough to merit the individual representation. The average fluxes of the three sections for each season are demonstrated (fig. 4). The geopotential difference at the sea surface between the south and north limit of the current is as large as one dynamic meter (table 1). The average total flux of the six sections is as much as $188 \text{ km}^3/\text{hr}$ ($52 \times 10^6 \text{ m}^3/\text{s}$). The value nearly equals the average value which was previously computed on the same assumption (MASUZAWA, 1954).

4.1. *Water characteristics in winter*

The range of salinity for a given temperature is less than 0.4‰ , actually 0.2‰ , over the whole range of temperature (3 to 21°C). In other words, the temperature-salinity curves of the Kuroshio water are fairly uniform over the total width of the Kuroshio. Almost all of characteristic classes on fig. 4a appear in each of the three different winters. The flux of the surface warm layer ($16\text{--}21^\circ\text{C}$) is about half the total flux. Two modes appear in the surface warm layer; the one centered at 16.5°C , 34.7‰ , the other at 18.5°C , 34.7‰ , but it is doubtful if they would be truly separate. The salinity maximum is in the isanosteric (uniform thermosteric-anomaly) layer of 300 to $350 \text{ c\ell/ton}$. According to our experience, the salinity of the Kuroshio south of Honshu very seldom exceeds 35‰ .

4.2. *Water characteristics in summer*

The flux of the warm water (more than 17°C) is half the total flux, and the temperature-salinity curves in the cold water (less than 17°C) are as uniform as they are in winter. The weak mode of flux can be seen in a few classes centered at 18°C , 34.75‰ . The range of salinity in the surface warm layer (more than 19°C) is remarkably wide, and the minimum salinity, 32.43‰ was observed at the sea surface near the current axis in 1953. It should be noted that the fairly large volume of fresh water is carried by the Kuroshio, that is, the flux of the less saline water (less than 34.5‰) in the surface warm layer is $26 \text{ km}^3/\text{hr}$ (15% of the total flux). However, all classes of water characteristics

of the surface warm layer in fig. 4b might not be found every summer or always in a summer.

4.3. Winter-summer differences

The total flux in winter is about $20 \text{ km}^3/\text{hr}$ larger than that in summer, but the difference may not be significant. In the range of $3\text{--}21^\circ\text{C}$, the fluxes of salinity-higher classes for any given temperatures are less in summer than in

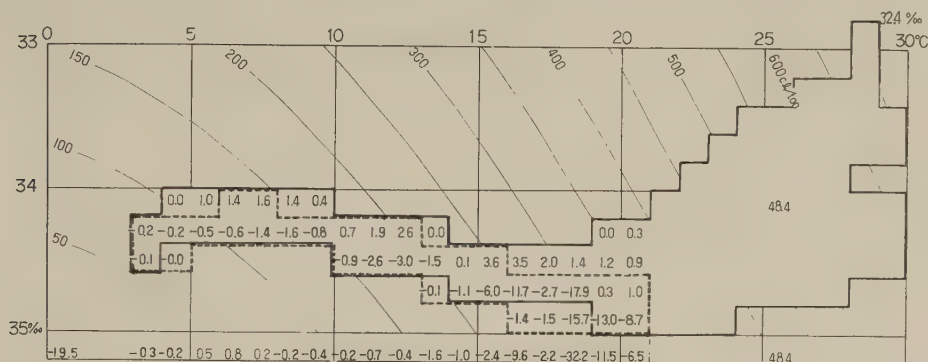


Fig. 5. The summer-winter difference of fluxes (km^3/hr) of the Kuroshio south of Cape Shiono-misaki. Positive sign means the summer-flux is larger.

winter, and the fluxes in salinity-lower classes increases from winter to summer. Namely, the salinity for a given temperature is rather lower in summer than in winter, not only in the surface layer but also in the subsurface layer. The flux of the top warm layer ($21\text{--}30^\circ\text{C}$) in summer is nearly equivalent to the decrease of the flux in the layer of $18\text{--}21^\circ\text{C}$ from winter to summer.

5. Differences between the North Equatorial Current and the Kuroshio

A similar representation of water characteristics of the North Equatorial Current has been made from the hydrographic sections taken in summer, 1956 along 140 and 150°E long. (MASUZAWA, 1964a). The total flux of the current is $176 \text{ km}^3/\text{hr}$, which nearly equals the value ($178 \text{ km}^3/\text{hr}$) of the Kuroshio south of Japan in summer.

Fig. 6 shows the difference of fluxes between the North Equatorial Current and the Kuroshio south of Japan in summer. The salinity-higher classes for any given temperatures in the North Equatorial Current disappear or decrease in flux in the Kuroshio. Particularly in the warm layer ($19\text{--}30^\circ\text{C}$), the most saline waters (say, more than 35‰) of the North Equatorial Current do not

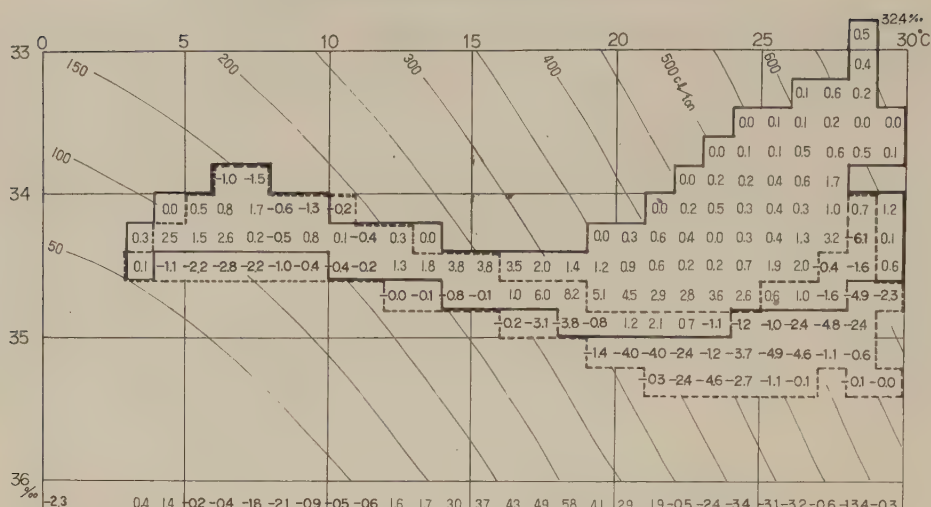


Fig. 6. The difference of fluxes (km^3/hr) between the Kuroshio and the North Equatorial Current in summer. Positive sign means the Kuroshio flux is larger.

appear in the Kuroshio. On the other hand, the fluxes in the classes of $17-21^\circ\text{C}$, $34.6-34.8\text{‰}$ increase outstandingly in the Kuroshio. In the intermediate layer (thermosteric anomaly, $100-150\text{ c}/\text{ton}$ or so), both the least saline water (less than 34.0‰) and the relatively salinity-higher water (more than 34.4‰) can not be found in the Kuroshio, and the fluxes in the classes ($34.0-34.4\text{‰}$) between these two waters increase in the Kuroshio.

In conclusion, the water characteristics of the Kuroshio may be changed considerably by various types of mixing and by air-sea processes in flowing along the western boundary region from Philippines to Japan. And also it should be noted that the fluxes of surface water characteristics of the Kuroshio south of Japan are fairly different between winter and summer.

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Preliminary Report of the Japanese Expedition of Deep-Sea, the Tenth Cruise, 1965 (JEDS-10)*

By

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Japan Meteorological Agency

(Received October 12, 1965)

1. *Introduction*

The Japanese Expedition of Deep-Sea, the Tenth Cruise (JEDS-10) which was planned by the Deep-Sea Research Committee (Chairman, K. WADATI), Japan Society for Promotion of Science, was carried out from April 22nd to May 9th, and from May 12th to 26th, 1965 on board the R/S Ryofu-Maru (Captain, T. SUENAGA) of the Japan Meteorological Agency.

The research stations were occupied at the neighbourhood of the northern part of the Izu-Ogasawara Submarine Ridge (Fig. 1).

In the former period, the original plan drew up various kinds of geophysical, geological, chemical, biological observation or sampling with the deep-sea winch. In the later period, the project aimed mainly to hydrographical researches of the deep-current, the bottom current and a few samplings of the plankton in the western side of the Ridge.

2. *The research members on board the Ryofu-Maru*

April 22-May 9

Japan Meteorological Agency: H. AKAMATSU (Phys. Ocean. Chief Scientist), Y. KAWARADA (Plankton), T. AKIYAMA (Chemistry), H. YOSHIMURA (Chemistry), T. NAKAI (Phys. Ocean.), A. SANO (Plankton), I. FUJIWARA (Phys. Ocean.), T. URASHIMA (Phys. Ocean.), T. YURA (Chemistry)

Meteorological Research Institute: Y. SUGIMURA (Chemistry)

* JEDS Contribution No. 65.

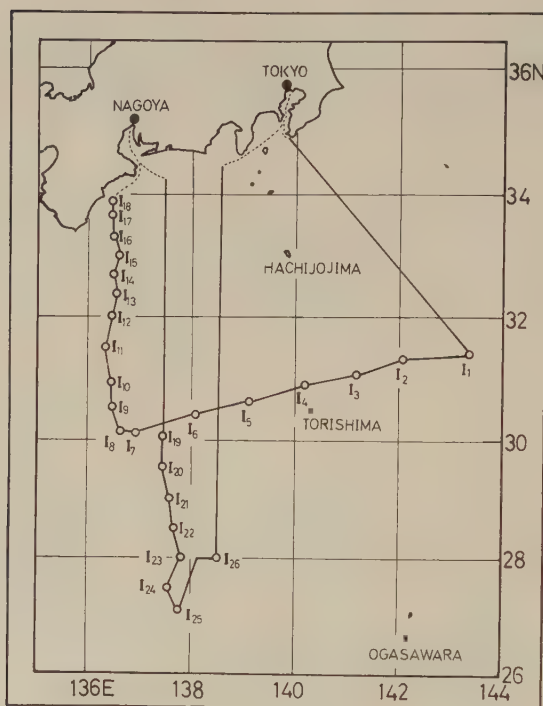


Fig. 1. The observational stations occupied by the Ryofu-Maru in the Japanese Expedition of Deep-Sea, the Tenth-Cruise (JEDS-10) in April 22 to May 26, 1965.

Hydrographic Office: T. SATO (Geology)

Tokyo University of Education: E. WADA (Chemistry), T. UCHIDA (Chemistry)

Tokyo University: E. HONDA (Geology)

Tokai University: M. OGURA (Phys. Ocean.)

Geological Survey of Japan: K. ONODERA (Geology), Y. SUMI (Geology)

Physical and Chemical Research Institute: S. WATANABE (Phys. Ocean.), M. KISHINO (Phys. Ocean.)

Japan Petroleum Exploration Co. Ltd.: T. ASAKAWA (Geology)

May 12-26

Japan Meteorological Agency: H. AKAMATSU (Chief Scientist), Y. KAWARADA, T. SAWARA (Phys. Ocean.), T. NAKAI, A. SANO, T. SAGI (Chemistry), I. FUJIWARA, T. URASHIMA, T. YURA, N. SATO (Phys.

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Ocean.)

Meteorological Research Institute: Y. SUGIMURA

Physical and Chemical Research Institute: G. OSHIBA (Phys. Ocean.),
M. KISHINO

3. *The items of observations or samplings*

- (1) Hydrographical cast from surface to bottom:
25 stations (I_1 — I_{26} , except I_8)
- (2) BT observations:
80 stations (Each hydro. station and every 3 hours)
- (3) GEK observations:
82 stations (Each hydro. station and every 3 hours)
- (4) Bottom-core sampling with gravity-corer:
2 stations (I_1 , I_5)
- (5) Bottom-core sampling with piston-corer:
4 stations (I_3 , I_4 , I_6 , I_7)
- (6) Dredging:
3 stations (I_4 , I_5 , I_8)
- (7) Deep water sampling for special analysis:
3 stations (I_2 , I_4 , I_{20})
- (8) Measurement of γ -ray activity:
2 stations (I_2 , I_7)
- (9) Sampling of deep water plankton:
3 stations (I_2 , I_4 , I_{23})
- (10) Measurement of bottom current:
3 stations (I_5 , I_6 , I_{23})
- (11) Measurement of deep water current:
1 station (I_{23})

4. *Results of the hydrographic cast*

We took twenty-five hydrographic stations along the latitude 31°N. , and two longitudes $136^\circ30'\text{E.}$, and $137^\circ30'\text{E.}$ (Fig. 1). In order to make the geostrophic computation exactly and to make the analysis of water properties especially in the deep water which will indicate the difference between the western side and the eastern side of the Ridge, the level observation was taken at intervals of 250 m, in the layers from 1000 m to bottom. The temperature measurement was

made very carefully with the thermometer which was calibrated repeatedly.

The salinity (reported as chlorinity) was measured with the inductive salinometer (Auto-Lab.) on land after the cruise. The pH measurement was carried out on board with the Toa-Denpa HM-5A pH meter reported as pH *in situ* (pHd).

Phosphate-P and silicate-Si were determined on board colorimetrically with the Tokyo-Koden D-7 recording electric photometer on the samples kept at the temperature of 20°C. by using a thermostat. Contents were calculated from each transmittance value referring to a calibration curve prepared previously at the same chemical elements other than the above, nitrate and carbonate and others, will be shortly reported.

The results are shown in Table 1.

5. Results of the other branches

Details of results in each branch shall be published in near future by other members and shall be collected in the "Reports of JEDS" published by the Deep Sea Research Committee, Japan Society for Promotion of Science.

Table 1. Results of serial observations in JEDS-10 April-May 1965.

St. I₁ (Ry 2174). 31°24'N, 143°25'E. April 25, 1965. Depth-5790 m.

Depth (m)	T (°C)	Cl (‰)	O ₂ (ml/l)	PO ₄ -P (μg-atoms/l)	Si (μg-atoms/l)	pHd
0	18.3	19.340	5.23	0.12	3	8.25
51	18.15	.322	5.15	0.15	3	8.31
104	17.97	.327	5.22	0.11	3	8.31
156	17.54	.312	5.11	0.23	3	8.24
208	17.34	.301	5.01	0.25	3	8.25
260	16.96	.287	4.88	0.28	4	8.27
311	16.51	.261	4.47	0.38	5	8.24
414	14.52	.155	4.21	0.79	9	8.18
516	12.23	.094	4.00	1.13	10	8.16
617	9.42	18.966	3.72	1.47	25	8.08
718	6.93	.911	2.95	1.87	40	7.96
822	5.20	.914	2.22	2.43	59	7.88
1033	3.83	.983	1.32	2.65	84	7.82
1296	2.96	19.048	0.89	3.00	103	7.75
1537	2.56	.088	0.96	3.01	115	7.74
1779	2.24	.126	1.25	3.07	120	7.78
2020	2.00	.147	1.65	2.97	123	7.77
2262	1.86	.164	2.12	2.87	121	7.78
2506	1.73	.176	2.40	2.86	118	7.86
2751	1.65	.188	2.64	2.66	118	7.88

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Depth (m)	T (°C)	Cl (‰)	O ₂ (ml/l)	PO ₄ -P (μg-atoms/l)	Si (μg-atoms/l)	pHd
2997	1.59	.191	2.82	2.67	117	7.92
3291	1.55	.194	3.01	2.64	115	7.91
3540	1.53	—	3.24	2.58	112	7.90
3789	1.51	—	3.32	2.58	112	7.89
4038	1.51	.204	3.39	2.44	109	7.90
4286	1.49	—	3.48	2.39	106	7.93
4535	1.48	.205	3.49	2.34	110	7.92
5034	1.53	.207	3.58	2.35	111	7.90
5533	1.59	.206	3.50	2.29	108	7.97

St. I₂ (Ry 2175). 31°18'N, 142°05'E. April 26, 1965. Depth-8850 m.

Depth (m)	T (°C)	Cl (‰)	O ₂ (ml/l)	PO ₄ -P (μg-atoms/l)	Si (μg-atoms/l)	pHd
0	18.1	19.298	5.39	0.12	3	8.23
54	18.03	.303	4.97	0.25	3	8.24
108	17.29	.316	5.13	0.23	3	8.24
160	17.23	.309	5.12	0.23	3	8.26
213	17.19	.308	5.07	0.26	3	8.27
266	17.04	.301	5.04	0.25	3	8.29
320	17.02	.300	5.02	0.28	4	8.23
425	15.71	.215	4.27	0.58	7	8.23
530	12.28	.077	3.83	1.04	13	8.10
633	9.55	18.982	3.62	1.43	22	8.07
736	7.24	.910	3.00	1.90	36	8.00
840	5.20	.873	2.33	2.19	54	7.87
1050	3.80	.970	1.13	2.70	85	7.79
1177	3.42	19.017	1.05	3.02	94	7.75
1397	2.93	.071	0.99	3.02	104	7.74
1621	2.48	.106	1.12	3.02	114	7.73
1848	2.18	.136	1.52	2.90	116	7.76
2080	1.98	.155	1.93	2.85	112	7.81
2313	1.85	.168	2.27	2.64	118	7.85
2767	1.70	.182	2.60	2.76	117	7.84
3005	1.61	.191	2.73	2.51	114	7.90
3252	1.55	.193	2.96	2.60	114	7.91
3382	1.56	—	—	2.64	114	7.92
3622	1.53	.201	3.17	2.58	113	7.92
3862	1.52	.200	3.25	2.51	105	7.92
4102	1.50	—	3.33	2.54	101	7.90
4343	1.48	.207	3.34	2.33	107	7.91
4827	1.54	.207	3.47	2.42	109	7.91
5314	1.58	.207	3.49	2.30	101	7.92
5801	1.62	.208	3.59	2.33	109	7.87
6287	1.68	.208	3.58	2.29	98	7.92
6776	1.76	.211	3.63	2.39	101	7.88

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St. I₃ (Ry 2176). 31°03'N, 141°14'E. April 28, 1965. Depth-3225 m.

Depth (m)	T (°C)	Cl (‰)	O ₂ (ml/l)	PO ₄ -P (μg-atoms/l)	Si (μg-atoms/l)	pHd
0	18.5	19.317	5.38	0.07	3	8.28
50	18.32	.342	5.16	0.11	5	.28
99	18.24	.345	5.07	0.21	5	.28
148	18.02	.329	5.02	0.22	5	.27
195	17.98	.324	5.00	0.18	5	.25
242	17.86	.324	4.97	0.25	4	.27
289	17.76	.322	5.01	0.29	5	.26
380	17.14	.309	5.07	0.32	6	.26
473	15.70	.219	4.43	0.57	9	.21
560	13.30	.109	4.08	0.87	12	.14
648	11.03	.022	3.76	1.25	64	.09
735	8.14	18.958	3.03	1.69	34	7.96
922	5.04	.988	1.99	2.51	66	.84
1088	4.00	19.017	1.57	2.81	79	.78
1319	2.97	.065	1.23	2.87	96	.73
1551	2.45	.105	1.31	2.97	106	.74
1786	2.18	.131	1.55	2.85	109	.74
2022	1.96	.153	2.00	2.87	109	.78
2263	1.82	.168	2.18	2.64	109	.83
2506	1.70	.178	2.51	2.74	109	.84
2749	1.64	.184	2.65	2.45	106	.86

St. I₄ (Ry 2177). 30°54'N, 140°14'E. April 28, 1965. Depth-1815 m.

Depth (m)	T (°C)	Cl (‰)	O ₂ (ml/l)	PO ₄ -P (μg-atoms/l)	Si (μg-atoms/l)	pHd
0	19.6	19.335	5.14	0.04	4	8.24
49	19.37	.341	5.14	0.07	2	.29
97	18.82	.324	5.01	0.17	4	.28
146	18.27	.312	4.90	0.21	3	.26
195	18.14	.323	4.85	0.23	2	.25
244	17.84	.308	4.68	0.26	4	.24
293	17.41	.289	4.55	0.33	4	.23
390	16.19	.238	4.34	0.48	5	.21
488	14.53	.161	4.11	0.90	6	.17
586	12.04	.059	3.86	1.12	12	.12
685	9.26	18.987	3.25	1.60	20	.10
784	6.80	.953	2.53	2.00	37	7.94
983	4.28	19.007	1.71	2.72	48	.86
1245	3.14	.063	1.37	2.95	88	.77
1508	2.56	.105	1.48	2.87	—	.76
1763	2.09	.136	1.54	2.80	—	.75

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St. I₅ (Ry 2178). 30°39'N, 139°08'E. April 30, 1965. Depth-1900 m.

Depth (m)	T (°C)	Cl (‰)	O ₂ (ml/l)	PO ₄ -P (μg-atoms/l)	Si (μg-atoms/l)	pHd
0	19.5	19.239	5.20	0.11	4	8.27
50	19.31	.270	5.12	0.14	6	.27
99	18.90	.314	4.91	0.15	6	.29
148	18.76	.324	4.90	0.15	6	.29
198	18.53	.324	4.81	0.15	4	.29
246	18.28	.323	4.98	0.19	3	.29
295	18.20	.321	4.91	0.20	5	.27
392	17.21	.281	4.51	0.32	7	.24
487	15.96	.232	4.39	0.52	8	.25
582	13.73	.124	4.03	0.87	11	.17
666	10.02	18.993	3.51	1.51	22	.06
771	7.44	.957	2.73	1.92	42	7.96
966	4.29	.994	1.59	2.77	74	.82
1220	3.24	19.063	1.45	2.97	93	.79
1470	2.47	.113	1.65	2.81	107	.81
1723	2.21	.135	1.98	2.79	106	.85

St. I₆ (Ry 2179). 30°25'N, 138°06'E. May 1, 1965. Depth-4105 m.

Depth (m)	T (°C)	Cl (‰)	O ₂ (ml/l)	PO ₄ -P (μg-atoms/l)	Si (μg-atoms/l)	pHd
0	19.6	19.334	5.19	0.07	3	8.27
54	18.92	.338	5.13	0.08	4	.28
109	18.27	.314	4.76	0.22	5	.26
163	17.91	.308	4.84	0.18	5	.26
218	17.62	.306	4.88	0.25	4	.26
268	17.22	.298	4.97	0.25	3	.24
319	16.86	.271	4.57	0.39	5	.23
421	15.32	.193	4.30	0.54	9	.20
519	13.66	.121	4.11	0.85	11	.16
618	10.74	.018	3.66	1.31	20	.10
719	8.32	18.963	3.10	1.73	33	.01
821	6.42	.956	2.34	2.17	52	7.91
1021	4.14	19.016	1.59	2.64	79	.79
1142	3.47	.051	1.47	2.87	92	.78
1335	2.86	.093	1.59	2.84	102	.80
1526	2.49	.118	1.85	2.81	105	.81
1719	2.22	.139	2.08	2.70	107	.85
1919	2.02	.158	2.37	2.76	108	.88
2130	1.86	.168	2.55	2.39	107	.90
2345	1.77	.176	2.80	2.66	108	.90
2563	1.69	.185	2.86	—	108	.92
2785	1.62	.192	3.12	2.50	107	.92
3012	1.57	.196	3.26	2.54	108	.91
3243	1.54	.198	3.31	2.52	104	.91
3477	1.54	.200	3.41	2.53	106	.91

St. I₇ (Ry 2180). 30°09'N, 136°56'E. May 2, 1965. Depth-3640 m.

Depth (m)	T (°C)	Cl (‰)	O ₂ (ml/l)	PO ₄ -P (μg-atoms/l)	Si (μg-atoms/l)	pHd
0	19.2	19.284	5.28	0.08	5	8.30
51	18.93	.293	5.26	0.07	4	.31
103	18.52	.315	4.96	0.14	5	.30
154	18.34	.319	4.93	0.15	4	.30
204	18.21	.328	4.99	0.14	4	.29
254	17.91	.316	4.85	0.18	3	.28
304	17.54	.309	4.95	0.20	4	.28
400	16.42	.249	4.46	0.47	6	.24
493	14.34	.156	4.14	0.80	10	.19
584	11.68	.045	3.91	1.12	16	.12
674	9.19	18.958	3.54	1.51	25	.05
765	6.98	.954	2.92	1.92	41	7.95
954	4.51	19.024	1.81	2.80	84	.82
980	4.31	.005	1.56	2.72	80	.83
1143	3.46	.054	1.56	2.73	90	.80
1312	2.91	.088	1.60	2.99	100	.80
1486	2.65	.108	1.77	2.73	103	.83
1660	2.41	.126	1.94	2.85	107	.85
1844	2.13	.145	2.01	2.57	108	.86
1868	2.11	.151	2.23	2.79	108	.86
2081	1.95	.165	2.42	2.60	107	.88
2299	1.81	.175	2.74	2.73	107	.90
2519	1.72	.184	2.92	2.70	107	.90
2738	1.63	.191	3.08	2.70	107	.91
2964	1.58	.193	3.23	2.70	106	.94

St. I₉ (Ry 2181). 30°30'N, 136°30'E. May 4, 1965. Depth-4320 m.

Depth (m)	T (°C)	Cl (‰)	O ₂ (ml/l)	PO ₄ -P (μg-atoms/l)	Si (μg-atoms/l)	pHd
0	19.3	19.302	5.07	—	4	—
10	19.21	.304	5.02	—	—	—
20	19.04	.302	5.15	—	—	—
30	19.02	.302	5.20	—	—	—
50	18.94	.302	5.28	0.06	4	—
75	18.74	.312	5.11	—	—	—
100	18.49	.325	5.03	0.18	4	—
163	18.32	.323	5.04	0.08	5	—
218	18.28	.321	4.97	0.10	4	—
272	18.10	.322	4.90	0.08	3	—
327	17.65	.298	4.61	0.22	5	—
438	16.30	.243	4.44	0.42	7	—
546	14.48	.156	4.15	0.66	10	—
655	11.14	.040	3.56	1.25	21	—
764	8.40	18.968	2.99	1.61	31	—

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Depth (m)	T (°C)	Cl (‰)	O ₂ (ml/l)	PO ₄ -P (μg-atoms/l)	Si (μg-atoms/l)	pHd
873	5.79	.957	2.14	2.32	55	—
1092	3.65	19.042	1.51	2.73	84	—
1284	3.11	.078	1.51	2.76	95	—
1525	2.60	.111	1.62	2.67	102	—
1767	2.24	.139	2.07	2.64	108	—
2008	2.00	—	2.38	2.61	106	—
2249	1.87	.172	2.60	2.51	108	—
2490	1.74	.179	2.78	2.38	107	—
2731	1.65	.189	3.05	2.51	102	—
2972	1.58	.193	3.10	2.23	102	—
3216	1.54	.198	3.27	2.38	105	—
3460	1.54	.200	3.40	2.42	102	—
3706	1.54	.201	3.39	2.58	102	—
3954	1.57	.203	3.47	2.50	100	—

St. I₁₀ (Ry 2182). 30°56'N, 136°30'E, May 4~5, 1965. Depth-4250 m.

Depth (m)	T (°C)	Cl (‰)	O ₂ (ml/l)	PO ₄ -P (μg-atoms/l)	Si (μg-atoms/l)	pHd
0	19.1	19.288	5.37	0.14	4	8.33
10	19.14	.288	5.19	—	—	.33
20	19.10	.286	5.17	—	—	.31
29	19.06	.286	5.14	—	—	.33
49	18.93	.297	5.14	0.11	5	.33
74	18.74	.311	4.98	—	—	.34
98	18.47	.318	4.99	0.19	6	.33
101	18.44	.319	4.98	0.16	4	.32
156	18.40	.322	4.97	0.14	4	.32
202	18.31	.319	5.01	0.17	3	.32
250	18.16	.321	4.86	0.22	5	.32
342	17.49	.295	4.52	0.35	5	.28
434	16.29	.247	4.39	0.50	6	.26
533	14.61	.166	4.14	0.72	7	.22
626	12.39	.077	3.85	1.05	13	.18
724	9.37	18.992	3.31	1.68	22	.10
921	5.25	.976	2.02	2.32	45	7.94
1236	3.29	19.064	1.55	2.95	68	.85
1461	2.76	.102	—	2.76	74	.85
1687	2.35	.127	1.90	2.83	83	.85
1913	2.12	.147	2.25	2.62	106	.88
2143	1.95	.161	2.52	2.48	99	.90
2377	1.78	.175	2.75	2.67	94	.95
2613	1.72	.182	2.96	2.58	99	.94
2852	1.65	.188	—	2.37	94	.98
3096	1.56	.197	3.31	2.44	98	.95
3342	1.54	—	—	—	—	—
3591	1.54	.199	3.40	2.52	100	.94
3842	1.54	.202	3.48	2.44	96	.94

St. I₁₁ (Ry 2183). 31°30'N, 136°24'E, May 5, 1965. Depth-4190 m.

Depth (m)	T (°C)	Cl (‰)	O ₂ (ml/l)	PO ₄ -P (μg-atoms/l)	Si (μg-atoms/l)	pHd
0	19.2	19.263	5.15	0.04	7	—
10	19.24	.265	5.10	—	—	—
20	19.23	.275	5.09	—	—	—
30	19.22	.266	5.10	—	—	—
50	19.22	.265	5.13	0.04	5	—
75	19.21	.268	5.06	—	—	—
101	19.17	.276	5.02	0.04	6	—
144	18.71	.319	—	0.10	4	—
188	18.62	.323	4.85	0.11	4	—
230	18.34	.324	4.91	0.11	3	—
273	18.21	.327	4.85	0.18	3	—
359	17.32	.286	4.46	0.35	5	—
440	16.33	.243	4.35	0.42	6	—
524	14.51	.158	4.07	0.67	9	—
611	12.19	.062	3.79	0.97	13	—
703	9.86	18.981	3.51	1.39	22	—
899	5.94	.961	2.28	2.13	53	—
1022	4.74	.988	1.67	2.55	70	—
1225	3.48	19.052	1.49	2.79	87	—
1434	2.86	.093	1.57	2.72	95	—
1654	2.47	.121	1.81	2.72	98	—
1875	2.16	.145	2.17	2.67	100	—
2105	1.94	.163	2.43	2.44	103	—
2340	1.80	.176	2.74	2.56	101	—
2580	1.72	.183	2.85	2.35	99	—
2805	1.61	.191	3.09	2.51	98	—
3037	1.55	—	—	—	—	—
3271	1.52	.197	3.30	2.44	101	—
3510	1.54	.202	3.41	2.41	96	—

St. I₁₂ (Ry 2184). 32°02'N, 136°31'E. May 5, 1965. Depth-4100 m.

Depth (m)	T (°C)	Cl (‰)	O ₂ (ml/l)	PO ₄ -P (μg-atoms/l)	Si (μg-atoms/l)	pHd
0	20.1	19.271	5.23	0.08	3	8.21
11	19.22	.269	5.20	—	—	.30
22	19.16	.265	5.19	—	—	.31
33	19.14	.263	5.19	—	—	.31
55	19.06	.271	5.11	0.13	4	.30
82	18.93	.298	5.01	—	—	.30
110	18.63	.320	4.94	0.14	3	.29
148	18.38	.325	4.94	0.14	5	.30
198	18.28	.325	4.93	0.14	2	.30
247	18.19	.332	4.94	0.15	3	.30
297	18.15	.327	4.88	0.18	3	.27
393	17.05	.276	4.47	0.34	6	.23
491	15.17	.189	4.18	0.54	8	.19
594	12.74	.091	3.91	0.88	13	.16
697	9.18	18.980	3.33	1.47	25	.04

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Depth (m)	T (°C)	Cl (‰)	O ₂ (ml/l)	PO ₄ -P (μg-atoms/l)	Si (μg-atoms/l)	pHd
799	7.12	.936	2.77	1.87	42	7.94
941	5.12	.962	1.89	2.42	68	.80
998	4.47	.984	1.74	2.47	75	.80
1151	3.80	19.032	1.45	2.63	85	.84
1357	3.12	.076	1.46	2.64	98	.80
1572	2.69	.106	1.68	2.64	100	.81
1795	2.34	.131	2.00	2.64	108	.83
2027	2.09	.151	2.29	2.54	107	.86
2260	1.92	.167	2.57	2.58	110	.86
2499	1.79	.179	2.72	2.38	113	.91
2742	1.69	.186	2.98	2.39	109	.91
2972	1.62	—	—	—	—	—
3208	1.55	.195	3.28	2.36	106	.92
3449	1.55	.198	3.45	2.44	108	.91

St. I₁₃ (Ry 2185). 32°24'N, 136°35'E. May 5~6, 1965. Depth-4190 m.

Depth (m)	T (°C)	Cl (‰)	O ₂ (ml/l)	PO ₄ -P (μg-atoms/l)	Si (μg-atoms/l)	pHd
0	19.1	19.278	5.14	0.04	4	—
10	19.08	.281	5.17	—	—	—
20	19.04	.280	5.12	—	—	—
30	19.02	.281	5.15	—	—	—
50	19.00	.281	5.13	0.06	4	—
75	18.96	.285	5.02	—	—	—
100	18.76	.311	4.97	0.08	6	—
170	18.40	.323	4.91	0.11	4	—
226	18.27	.325	4.89	0.11	4	—
282	18.00	.317	4.75	0.18	4	—
338	17.32	.286	4.48	0.28	5	—
450	15.80	.218	4.31	0.51	7	—
559	12.72	.088	3.72	0.98	15	—
668	9.78	18.994	3.38	1.46	25	—
777	7.24	.931	2.78	1.90	39	—
889	5.64	.947	2.08	2.35	58	—
1118	3.78	19.035	1.53	2.64	83	—
1296	3.11	.076	1.53	2.70	94	—
1546	2.65	.108	1.72	2.68	101	—
1794	2.28	.136	1.99	2.59	104	—
2044	2.03	.157	2.31	2.59	106	—
2292	1.86	.170	2.69	2.53	104	—
2539	1.74	.182	2.82	2.36	103	—
2788	1.64	.187	3.05	2.52	103	—
3035	1.58	.196	3.09	2.28	105	—
3283	1.54	.198	3.29	2.35	101	—
3531	1.54	—	—	—	—	—
3779	1.54	.202	(3.33)	2.41	101	—
4026	1.58	.200	—	2.38	100	—

St. I₁₄ (Ry 2186). 32°40'N, 136°31'E. May 6, 1965. Depth-4500 m.

Depth (m)	T (°C)	Cl (‰)	O ₂ (ml/l)	PO ₄ -P (μg-atoms/l)	Si (μg-atoms/l)	pHd
0	19.7	19.265	5.16	0.07	2	—
10	19.82	.267	5.11	—	—	—
20	19.67	.262	5.15	—	—	—
30	19.65	.259	5.11	—	—	—
50	19.53	.269	5.06	0.08	4	—
75	19.19	.298	4.96	—	—	—
100	18.58	.323	5.04	0.04	5	—
177	18.33	.322	5.01	0.10	5	—
235	18.19	.328	4.91	0.13	3	—
294	17.76	.304	4.64	0.22	4	—
353	16.79	.272	4.52	0.38	6	—
471	14.36	.153	4.05	0.72	10	—
590	10.26	.005	3.47	1.33	22	—
699	7.34	18.937	2.86	1.89	37	—
809	5.68	.933	2.22	2.20	55	—
918	4.48	.985	1.59	2.60	70	—
1138	3.44	19.056	1.50	2.66	81	—
1250	3.17	.082	1.46	2.66	99	—
1480	2.66	.109	1.64	2.54	95	—
1711	2.31	.136	1.96	2.45	100	—
1944	2.10	.152	2.21	2.38	105	—
2180	1.94	.166	2.50	2.52	102	—
2418	1.81	.177	2.68	2.24	103	—
2659	1.70	.187	2.97	2.29	101	—
2903	1.62	.194	3.03	2.13	100	—
3147	1.56	.199	3.28	2.29	98	—
3393	1.55	.203	3.34	2.33	97	—
3642	1.55	.205	3.32	2.32	—	—
3891	1.58	.204	3.43	2.16	93	—
4141	1.58	.213	3.43	2.17	93	—

St. I₁₅ (Ry 2187). 33°01'N, 136°39'E. May 6, 1965. Depth-3390 m.

Depth (m)	T (°C)	Cl (‰)	O ₂ (ml/l)	PO ₄ -P (μg-atoms/l)	Si (μg-atoms/l)	pHd
0	22.1	19.179	4.86	0.07	3	—
11	22.33	.177	4.80	—	—	—
22	22.27	.176	4.82	—	—	—
33	22.27	.176	4.81	—	—	—
54	22.20	.180	4.77	0.08	4	—
81	21.31	.249	4.54	—	—	—
108	20.33	.306	4.87	0.07	5	—
141	19.28	.320	4.79	0.10	5	—
184	18.78	.329	—	0.04	3	—
227	18.03	.304	4.53	0.18	5	—

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Depth (m)	T (°C)	Cl (‰)	O ₂ (ml/l)	PO ₄ -P (μg-atoms/l)	Si (μg-atoms/l)	pHd
270	17.13	.277	4.49	0.31	5	—
352	15.50	.197	4.28	0.52	10	—
434	12.57	.109	3.60	1.09	17	—
508	9.87	.039	3.06	1.47	29	—
579	7.26	18.933	3.00	1.77	38	—
640	6.51	.937	2.57	2.25	53	—
650	6.76	.983	2.37	2.14	50	—
737	5.85	.978	2.02	2.36	57	—
795	4.82	19.005	1.83	2.32	65	—
815	4.70	18.987	1.65	2.59	71	—
884	4.05	19.017	1.46	2.72	75	—
974	3.75	.039	1.47	2.77	85	—
1051	3.40	.058	1.41	2.69	86	—
1149	3.18	.076	1.49	2.80	92	—
1264	2.94	.091	1.51	2.59	95	—

St. I₁₆ (Ry 2188). 33°20'N, 136°32'E. May 6, 1965. Depth-1975 m.

Depth (m)	T (°C)	Cl (‰)	O ₂ (ml/l)	PO ₄ -P (μg-atoms/l)	Si (μg-atoms/l)	pHd
0	20.7	19.133	5.18	0.02	5	—
10	20.73	.136	5.06	—	—	—
20	20.70	.144	5.02	—	—	—
30	20.69	.145	4.93	—	—	—
50	20.02	.188	4.56	0.11	3	—
75	18.16	.234	4.90	—	—	—
100	17.66	.261	4.25	0.32	8	—
151	15.14	.190	4.13	0.60	10	—
192	14.30	.163	3.98	0.71	13	—
231	13.55	.138	3.80	0.85	14	—
264	13.08	.118	3.69	0.90	15	—
323	10.93	.059	3.22	1.24	23	—
379	9.70	.026	2.98	1.45	29	—
431	8.07	18.979	2.79	1.77	39	—
478	7.02	.975	2.41	2.00	47	—
523	6.41	.983	2.23	2.14	54	—
604	5.10	.990	1.83	2.37	67	—
713	4.13	19.016	1.52	2.63	79	—
861	3.63	.041	1.43	2.66	87	—
1018	3.25	.065	1.45	2.64	90	—
1194	2.93	.088	1.53	2.66	94	—

St. I₁₇ (Ry 2189). 33°40'N, 136°29'E. May 6~7, 1965. Depth-2040 m.

Depth (m)	T (°C)	Cl (‰)	O ₂ (ml/l)	PO ₄ -P (μg-atoms/l)	Si (μg-atoms/l)	pHd
0	17.6	19.104	5.49	0.07	6	—
10	17.04	.152	5.13	—	—	—
20	16.02	.137	5.11	—	—	—
30	15.50	.158	4.94	—	—	—
50	14.73	.157	4.95	0.50	10	—
75	13.50	.150	4.82	—	—	—
100	13.09	.143	4.56	0.76	15	—
150	11.46	.099	4.11	1.13	21	—
200	9.35	.021	3.04	1.61	36	—
250	8.61	18.991	2.87	1.75	37	—
300	7.92	19.000	2.70	1.84	43	—
400	6.66	18.972	2.34	2.04	55	—
505	5.38	.979	1.91	2.28	69	—
610	4.49	19.001	1.64	2.47	82	—
715	4.01	.025	1.54	2.51	90	—
820	3.67	.043	1.53	2.59	97	—
1030	3.25	.068	1.54	2.57	104	—
1250	2.86	.094	1.68	2.66	112	—
1500	2.50	.121	1.95	2.60	114	—
1750	2.22	.143	2.15	2.48	111	—
2000	1.98	.161	2.45	2.37	116	—

St. I₁₈ (Ry 2190). 33°54'N, 136°29'E. May 7, 1965. Depth-1260 m.

Depth (m)	T (°C)	Cl (‰)	O ₂ (ml/l)	PO ₄ -P (μg-atoms/l)	Si (μg-atoms/l)	pHd
0	14.9	18.882	5.90	1.39	4	—
10	14.92	.955	5.82	—	—	—
20	14.68	.970	5.82	—	—	—
30	14.58	19.032	5.54	—	—	—
50	14.80	.093	5.42	0.41	6	—
75	14.24	.137	5.12	—	—	—
100	13.91	.141	4.94	0.63	11	—
151	12.40	.125	4.63	0.93	13	—
200	10.53	.052	4.14	1.36	23	—
248	9.03	.006	3.64	1.57	30	—
297	8.43	18.998	2.97	1.83	32	—
392	7.04	.974	2.59	1.98	41	—
487	5.93	.973	2.32	2.26	55	—
582	4.84	.991	1.91	2.44	68	—
677	4.32	19.013	1.77	2.51	70	—
774	3.93	.029	1.71	2.66	75	—
970	3.37	.061	1.65	2.66	80	—

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St. I₁₉ Ry 2191). 30°03'N, 137°28'E. May 13~14, 1965. Depth-4525 m.

Depth (m)	T (°C)	Cl (‰)	O ₂ (ml/l)	PO ₄ -P (μg-atoms/l)	Si (μg-atoms/l)	pHd
0	21.0	19.286	5.24	0	1	—
10	20.90	.288	5.18	0	2	—
20	20.90	.286	5.12	0	2	—
29	20.92	.293	5.13	0	1	—
49	20.11	.341	5.12	0	1	—
74	19.69	.341	5.05	0	1	—
98	19.28	.336	4.89	0.03	2	—
140	18.88	.336	4.99	0.02	2	—
184	18.38	.323	4.76	0.07	2	—
229	17.80	.305	4.65	0.14	2	—
273	17.24	.282	4.52	0.19	3	—
354	16.20	.240	4.45	0.36	4	—
434	14.82	.175	4.19	0.50	6	—
513	12.75	.085	4.02	0.71	9	—
591	10.38	18.996	3.71	0.93	17	—
671	8.28	.935	3.29	1.28	28	—
839	5.11	.913	2.10	2.02	54	—
1060	3.84	19.015	1.39	2.51	84	—
1268	3.08	.065	1.49	2.59	98	—
1471	2.64	.104	1.60	2.50	101	—
1673	2.36	.126	1.85	2.59	106	—
1876	2.11	.141	2.02	2.47	108	—
2089	1.94	.167	2.48	2.41	107	—
2307	1.82	.176	2.71	2.40	104	—
2528	1.74	.185	2.93	2.12	101	—
2751	1.65	.189	3.05	2.35	102	—
2976	1.60	.194	3.20	2.35	101	—
3206	1.55	.198	3.27	2.35	99	—
3440	1.55	.203	3.42	2.36	99	—
3681	1.53	.205	3.52	2.29	99	—

St. I₂₀ (Ry 2192). 29°33'N, 137°28'E. May 14, 1965. Depth-4265 m.

Depth (m)	T (°C)	Cl (‰)	O ₂ (ml/l)	PO ₄ -P (μg-atoms/l)	Si (μg-atoms/l)	pHd
0	20.7	19.246	5.21	tr	4	—
10	20.32	.248	5.14	—	—	—
20	20.12	.294	5.20	—	—	—
30	19.59	.326	5.25	—	—	—
50	19.16	.322	5.18	0.03	6	—
75	18.63	.326	5.24	—	—	—
100	18.52	.327	5.12	0.08	6	—
149	18.22	.324	4.97	0.14	5	—
199	17.94	.313	4.84	0.20	5	—
249	17.14	.278	4.54	0.35	6	—
298	16.73	.261	4.52	0.41	6	—
399	15.43	.199	4.31	0.48	8	—
499	12.72	.088	3.81	1.01	15	—
599	9.65	18.978	3.48	1.40	23	—
699	7.11	.916	2.92	1.89	42	—
800	5.57	.931	2.20	2.22	60	—
995	4.12	19.015	1.54	2.58	83	—
1233	3.08	.060	1.22	2.79	101	—
1476	2.61	.107	1.70	2.66	106	—

St. I₂₁ (Ry 2193). 29°00'N, 137°36'E. May 15, 1965. Depth—m.

Depth (m)	T (°C)	Cl (‰)	O ₂ (ml/l)	PO ₄ -P (μg-atoms/l)	Si (μg-atoms/l)	pHd
0	20.2	19.229	5.22	0	3	—
11	20.16	.235	5.13	—	—	—
22	20.13	.235	5.21	—	—	—
33	20.00	.243	5.27	—	—	—
54	19.30	.313	5.34	0	3	—
82	18.48	.301	5.04	—	—	—
109	18.31	.311	4.95	0.10	4	—
152	18.11	.313	4.87	0.18	4	—
203	17.81	.300	4.73	0.21	4	—
255	17.18	.275	4.55	0.32	5	—
309	16.66	.254	4.52	0.38	6	—
414	14.87	.171	4.20	0.55	9	—
527	12.41	.073	3.92	0.93	14	—
635	9.08	19.964	3.61	1.47	25	—
655	8.32	.929	3.36	1.53	29	—
765	6.20	.899	2.63	2.04	47	—
976	4.25	.997	1.64	2.51	74	—
1227	3.14	19.065	1.41	2.83	94	—
1491	2.55	.101	—	—	—	—

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St. I₂₂ (Ry 2194). 28°32'N, 137°44'E. May 16, 1965. Depth-4805 m.

Depth (m)	T (°C)	Cl (‰)	O ₂ (ml/l)	PO ₄ -P (μg-atoms/l)	Si (μg-atoms/l)	pHd
0	22.5	19.221	5.04	0	3	—
10	22.33	.222	5.15	—	—	—
19	22.26	.227	5.29	—	—	—
30	22.26	.239	5.30	—	—	—
50	21.24	.327	5.50	0	3	—
75	20.33	.331	5.10	—	—	—
100	19.20	.321	5.45	0.02	4	—
158	18.49	.324	5.19	0.10	4	—
205	18.37	.318	5.08	0.11	4	—
253	18.20	.317	5.09	0.18	4	—
301	17.93	.307	4.94	0.18	4	—
397	16.28	.236	4.44	0.39	6	—
496	14.50	.156	4.12	0.66	10	—
598	10.90	.013	3.87	1.15	18	—
692	8.15	18.938	3.32	1.48	30	—
789	6.11	.898	2.71	2.12	47	—
984	4.07	.956	1.44	2.52	76	—
1075	3.68	19.001	1.10	2.73	88	—
1252	3.11	.054	1.25	2.76	98	—
1431	2.65	.097	1.66	2.62	98	—
1449	2.65	.102	1.59	2.73	103	—
1596	2.45	.117	1.71	2.54	103	—
1612	2.42	.116	1.73	2.67	102	—
1756	2.23	.133	1.97	2.72	106	—
1918	2.09	.148	2.15	2.37	109	—
2091	1.94	.161	2.51	2.72	110	—
2258	1.87	—	—	—	—	—
2432	1.80	.172	2.80	2.56	109	—
2619	1.76	.180	2.95	2.56	109	—
2820	1.67	.192	3.12	2.47	109	—

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St. I₂₅ (Ry 2195). 27°08'N, 137°49'E. May 17, 1965. Depth-4660 m.

Depth (m)	T (°C)	Cl (‰)	O ₂ (ml/l)	PO ₄ -P (μg-atoms/l)	Si (μg-atoms/l)	pHd
0	24.4	19.205	4.83	—	8	—
10	24.33	.210	4.79	—	—	—
20	24.31	.215	4.86	—	—	—
30	24.29	.230	4.87	—	—	—
51	21.04	.319	5.30	—	5	—
77	20.44	.343	5.33	—	—	—
102	19.88	.341	4.98	—	6	—
169	18.66	.322	4.82	—	6	—
225	17.59	.290	4.81	—	6	—
281	16.86	.267	4.60	—	—	—
337	16.31	.244	4.59	—	5	—
449	14.41	.149	4.29	—	9	—
560	11.58	.029	4.06	—	15	—
671	8.34	18.915	3.63	—	27	—
782	6.28	.884	2.78	—	42	—
893	4.87	.925	1.92	—	60	—
1115	3.70	19.009	1.37	—	81	—
1503	2.64	.096	1.46	—	99	—
1734	2.35	.127	1.83	—	104	—
1964	2.06	.152	2.24	—	105	—
2195	1.87	.168	2.59	—	106	—
2425	1.78	.178	2.83	—	105	—
2656	1.71	.183	2.95	—	101	—
2890	1.65	.188	3.11	—	103	—
3124	1.62	.193	3.17	—	114	—
3361	1.58	.195	3.28	—	101	—
3601	1.55	.198	3.42	—	96	—
3843	1.54	.199	3.48	—	103	—
4087	1.58	.201	3.53	—	107	—
4333	1.58	.203	3.52	—	104	—

St. I₂₄ (Ry 2196). 27°31'N, 137°34'E. May 17, 1965. Depth-4860 m.

Depth (m)	T (°C)	Cl (‰)	O ₂ (ml/l)	PO ₄ -P (μg-atoms/l)	Si (μg-atoms/l)	pHd
0	23.8	19.269	4.86	0	3	—
10	23.57	.272	4.84	—	—	—
20	22.08	.276	5.00	—	—	—
30	21.38	.300	5.09	—	—	—
52	20.14	.333	5.23	0	4	—
79	18.98	.336	5.02	—	—	—
106	18.66	.330	5.06	0.05	4	—
161	18.07	.313	4.77	0.17	5	—
215	17.70	.304	4.78	0.17	4	—
254	17.30	.290	4.70	0.21	5	—

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Depth (m)	T (°C)	Cl (‰)	O ₂ (ml/l)	PO ₄ -P (μg-atoms/l)	Si (μg-atoms/l)	pHd
356	16.14	.217	4.40	0.44	6	—
482	14.05	.124	4.19	0.69	10	—
598	11.02	.001	4.01	1.13	12	—
701	8.20	18.927	3.26	1.54	30	—
805	6.27	.904	2.54	2.11	47	—
1013	4.14	.973	1.48	2.51	77	—
1287	3.09	19.058	1.26	2.73	98	—
1562	2.42	.111	1.57	2.66	110	—

St. I₂₃ (Ry 2197). 27°59'N, 137°51'E. May 18, 1965. Depth-4735 m.

Depth (m)	T (°C)	Cl (‰)	O ₂ (ml/l)	PO ₄ -P (μg-atoms/l)	Si (μg-atoms/l)	pHd
0	23.5	19.260	4.71	0	2	—
10	23.41	.263	4.77	—	—	—
20	22.98	.274	4.99	—	—	—
30	22.42	.308	5.10	—	—	—
50	21.43	.328	4.93	0	2	—
75	20.10	.334	4.83	—	—	—
100	19.56	.337	4.86	0.07	3	—
163	18.78	.338	4.99	0.08	3	—
217	18.11	.324	4.93	0.08	3	—
271	17.79	.319	4.99	0.11	3	—
326	17.09	.275	4.58	0.28	4	—
430	15.31	.195	4.41	0.50	6	—
532	12.53	.066	4.00	0.88	11	—
635	9.42	18.961	3.57	1.42	21	—
738	7.16	.904	3.03	1.74	34	—
841	5.43	.906	2.27	2.34	53	—
1047	3.90	.982	1.37	2.61	75	—
1248	3.13	19.052	1.25	2.79	91	—
1499	2.63	.101	1.58	2.37	98	—
1748	2.22	.134	1.95	2.51	100	—
1996	2.00	.155	2.38	2.48	100	—
2243	1.86	.169	2.67	2.44	110	—
3491	1.76	.178	2.88	2.27	96	—
2737	1.68	.186	3.07	2.36	96	—
2983	1.64	.191	3.11	2.25	98	—
3231	1.57	.196	3.37	2.36	98	—
3476	1.56	.199	3.39	2.35	96	—
3993	1.56	.200	3.44	2.14	90	—
4242	1.61	.202	3.48	2.28	96	—
4490	1.60	.203	3.50	2.25	96	—

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St. I₂₆ (Ry 2198). 28°00'N, 138°32'E. May 22, 1965. Depth-4390 m.

Depth (m)	T (°C)	Cl (‰)	O ₂ (ml/l)	PO ₄ -P (μg-atoms/l)	Si (μg-atoms/l)	pHd
0	24.4	19.210	4.96	0	2	—
10	24.11	.243	4.84	—	—	—
20	22.46	.263	5.06	—	—	—
30	22.06	.309	5.10	—	—	—
50	20.96	.343	5.05	0	2	—
76	20.13	.339	4.97	—	—	—
102	19.48	.334	4.79	0.11	3	—
164	18.52	.324	5.00	0.11	3	—
221	18.24	.323	4.91	0.14	3	—
279	17.75	.300	4.69	0.22	4	—
337	17.10	.278	4.63	0.28	4	—
453	15.34	.195	4.31	0.57	7	—
569	12.45	.062	4.06	0.85	12	—
685	9.06	18.946	3.48	1.54	24	—
793	6.37	.895	2.78	2.00	47	—
901	5.02	.913	1.98	2.44	61	—
1118	3.58	19.009	1.22	2.73	87	—
1249	3.27	.045	1.26	2.88	96	—
1501	2.62	.094	1.37	2.79	106	—
1751	2.26	.128	1.77	2.54	108	—
2000	2.02	.152	2.26	2.54	108	—
2249	1.85	.168	2.60	2.53	109	—
2498	1.77	.177	2.80	2.44	106	—
2747	1.68	.184	3.01	2.44	103	—
2995	1.64	.189	3.15	2.20	104	—
3245	1.59	.193	3.27	2.41	105	—

On an Instrument for *in situ* Measurement of γ -Ray Activity in the Deep Water of the Ocean*

By

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(Received October 27, 1965)

Abstract: For the purpose of monitoring the artificial radioactive contamination in the ocean, a special instrument has been built for *in situ* measurement of γ -ray activity in deep water.

Results of observations in the adjacent seas of Japan and of calibration tests are reported.

1. *Instrument and procedures in the field*

In recent years, with the development of nuclear-energy exploration, contamination with artificial radioactivity has become a serious problem.

In order to monitor the artificial contamination in the ocean a special instrument has been built for *in situ* measurement of γ -ray activities from the surface to a depth of 3000 m without sampling water.

As shown in Fig. 1, the instrument consists of main five parts, namely, a detector, a counting circuit, a time-notching circuit, a time and a pulse recorder, and an electric power supply. Each part is all transistorized and made compact, and the whole parts are put together in a steel container.

Details of each part are as follows:

a) Container material: Steel pipe without patchwork.

Size: Length —2 m.

Diameter —110 mm.

Thickness—10 mm except around the detector where it is 5 mm thick for obtaining high level in measurement sensitivity.

Maximum pressure durable: 450 Kg/cm².

Total weight: 50 Kg.

* JEDS Contribution No. 67

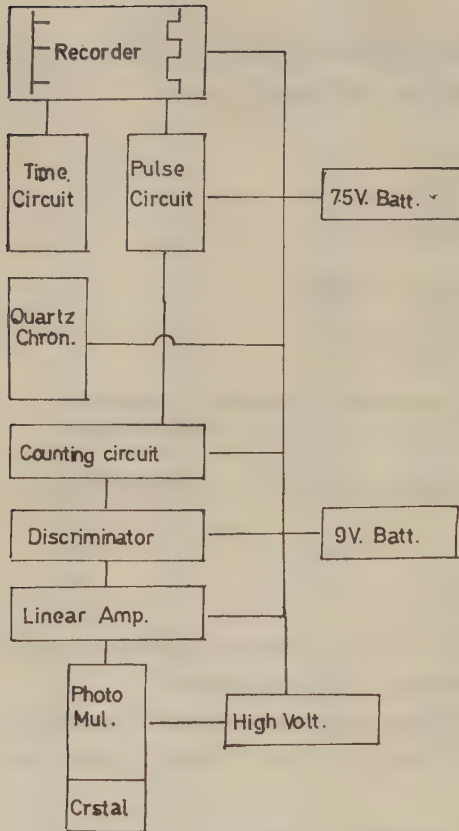


Fig. 1. Schematic diagram of the instrument.

- b) Detector: NaI (Tl) crystal, 1.5 inch ϕ × 1.5 inch height. The energy level below 60 KeV is cut off by the discriminator.
- c) Counting circuit: Counting rate is scale down to 1/16 or 1/32.
- d) Time checking: Time mark is put every minute by use of a quartz chronometer.
- e) Recorder: Stirrus paper is driven by a micromotor at a speed of 1 cm/min. Both pulse and time marks are plotted simultaneously.
- f) Energy supply: Alkaline dry battery, Maxwell UM-2 (1.5V.), by which constant voltage can be maintain for about 30 hours between 0° and 40°C.

Prior to the present measuring system, another system converting a pulse to an ultrasonic wave, and receiving it with a hydrophone and recording it on board was tested,

and found unsuitable for a practical field use. As an intensity of ultrasonic wave received with the hydrophone diminishes with depth, sensitivity of the receiver must be adjusted to higher level when the instrument is lowered down deeper. But raising sensitivity strengthens a noise intensity. In measuring activities in deep water it was difficult to judge whether a signal received was due to noise or not. In order to avoid the disturbance by the noise the present recording system has been adopted in spite of an inconvenience, that is unawareness of results until the instrument has been hauled onto the board. In the present system, activities at depths can be read from the record.

Procedures at field measurement are as follows:

- 1) Fix batteries and a recording paper in their places with screw nuts.
- 2) Checking on board a normal operation of the instrument for a few minutes

before lowering the instrument down. Put on switches and record that time. Close the container tightly with "O-ring" and screw nuts.

- 3) Connect the instrument with an end of a sounding wire and lower it down into the sea. At each depth where observation is to be done, hold the instrument at the depth for longer than 5 minutes.
- 4) After hauling the instrument onto the board, open the container and remove the recording paper.
- 5) An activity at each depth is known by counting numbers of pulse marks on the recording paper.

Here, caution must be paid that humidity inside the container should be kept low. Otherwise, spurious counting is observed.

2. Calibration

In order to estimate the quantity of a dissolved matter from its radioactivity, it is necessary to establish a numerical relation between the quantity of each dissolved nuclide and activity detected by the present instrument. Activities of a certain known γ -emmitter dissolved in sea water may be calculated by the following equation. A model assumed for the calculation is shown in Fig. 2.

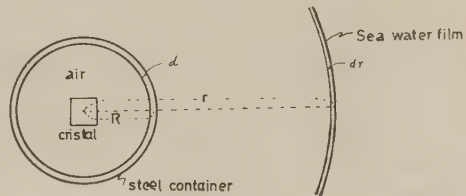


Fig. 2. A model for the theoretical calculation of γ -ray activity.

$$I = E \int_R^\infty N \cdot 4\pi r^2 dr \frac{S}{4\pi r^2} \cdot e^{-\mu_w(r-R)} \cdot e^{-\mu_F \cdot d}$$

$$= E \frac{N \cdot S}{\mu_w} \cdot e^{-\mu_F \cdot d}$$

where

I : activity (c.p.m.)

E : counting efficiency

N : $\bar{c} \times \frac{6.02 \times 10^{23}}{M} \times \lambda$, disintegration numbers of a nuclide in unit volume of sea water in unit time

S : average cross section of NaI (Tl) crystal (cm^2)

μ_w, μ_F : linear absorption coefficient of sea water and steel respectively (cm^{-1})

d : thickness of the container (cm)

According to the previous equation, the activity of K-40 uniformly distributed in the sea is estimated as shown below.

$$I = 6 \text{ c.p.m.}$$

On the other hand, a calibration in a water tank was made empirically. The instrument was placed at the center of the water tank, $3 \times 4 \times 2 \text{ m}$ in size. Preliminarily natural back ground activity was measured every ten minute after the tank was filled with water free from potassium. From the measurement $100.5 \pm 0.9 \text{ c.p.m.}$ was obtained as back ground. A reagent grade of potassium chloride was added stepwise. A counting rate was measured at each addition. As shown in Fig. 3, a relation between a counting rate and a corresponding

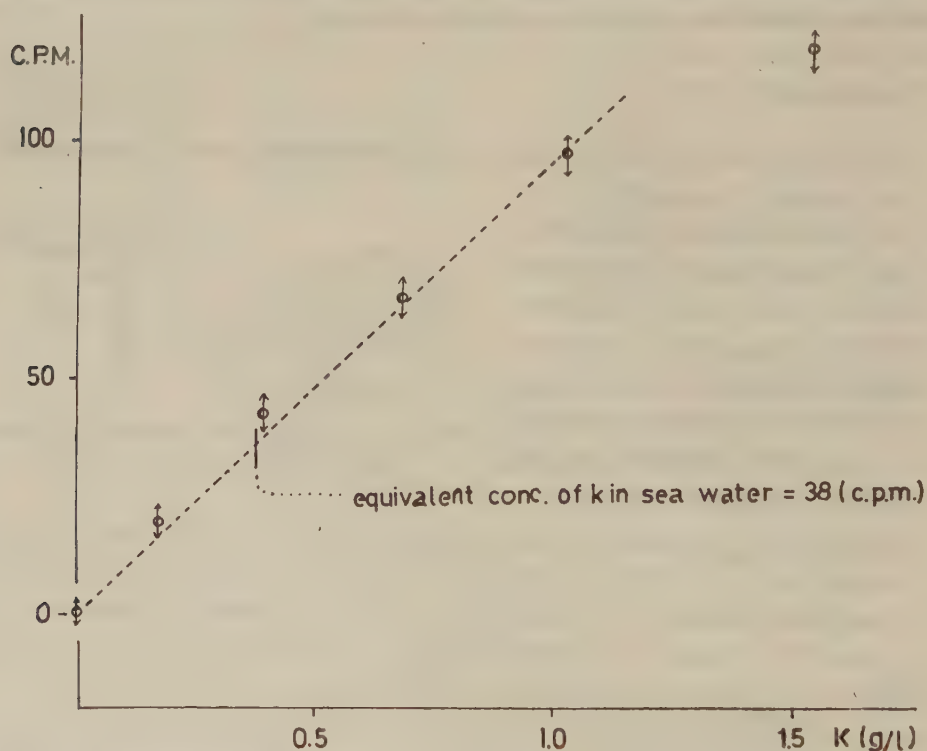


Fig. 3. A calibration result between a counting rate (c. p. m.) and a potassium concentration dissolved in water.

potassium concentration was almost linear except the region in high concentration which is five times more concentrated than in sea water. From Fig. 3 activity of potassium chloride solution which are equivalent in potassium concentration to sea water with a chlorinity of 19.00‰ can be read as 38 c.p.m.. As shown in

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the following clause, the mean value of activity in deep water detected by the present instrument is nearly 60 c.p.m., so the residual 22 c.p.m., that is, (60-38) c.p.m. may probably be derived from radioactive nuclides other than potassium.

The discrepancy between results of theoretical calculation and of empirical calibration is not clear at present, but is now being pursued.

3. Results and discussions

Results obtained are tabulated in Table 1. Vertical distribution of activity is shown in Fig. 4. In the layer shallower than 100 m, activities are relatively higher and fluctuate remarkably from station to station than in the deeper one.

Table 1. Results measured of γ -ray activities in the adjacent seas of Japan.

Station	A	B	C	D	E	C'	F	G	C''	mean $\pm\sigma$
Date	Nov. 17 1962	May 26 '63	Jul. 24 '63	Sep. 28 '63	May 18 '64	Feb. 22 '65	Apr. 25 '65	May 1 '65	May 25 '65	
Location Lat.	N 33-00	40-51	34-43	38-06	42-08	34-37	31-18	30-09	34-40	
Long.	E 138-00	144-48	139-54	149-49	155-56	140-00	142-05	136-56	140-25	
Depth m										
0	212	144	328	128			151	147	120	175.7 $\pm$ 68.0
2			128	91		110	126	115		114.0 $\pm$ 13.3
5			109	91	92	96	95	102	98	97.6 $\pm$ 5.8
10	90	96	98	86	83	82	90	80	82	87.4 $\pm$ 6.1
15			88	83	75					82.0 $\pm$ 5.4
20	85	80	80	82	66	76	73	70		76.5 $\pm$ 6.0
30	80	72	78	75	71	69	69	67		72.6 $\pm$ 4.3
40	74	64	78	75	64					71.0 $\pm$ 1.9
50	68	64	70	75	62	59	59	66	66	65.4 $\pm$ 5.0
100	70	64	67	64	68	58	68	62	69	65.6 $\pm$ 3.7
200					56	56	59			57.0 $\pm$ 1.4
500	62	64	64	61	58	61	57	63	61	61.2 $\pm$ 2.3
1,000	63	60	61	61	52	60	59	59	59	59.3 $\pm$ 2.9
1,500	63	64	59	58	57				59	60.0 $\pm$ 2.8
2,000	62	59	61	56	63	57	56	62	59	59.4 $\pm$ 2.6
2,500	60	59	60	64	59	71			58	61.6 $\pm$ 4.1
3,000	60	59		64	62		58	56		59.8 $\pm$ 2.6
3,500	61	59		64						61.3 $\pm$ 2.1

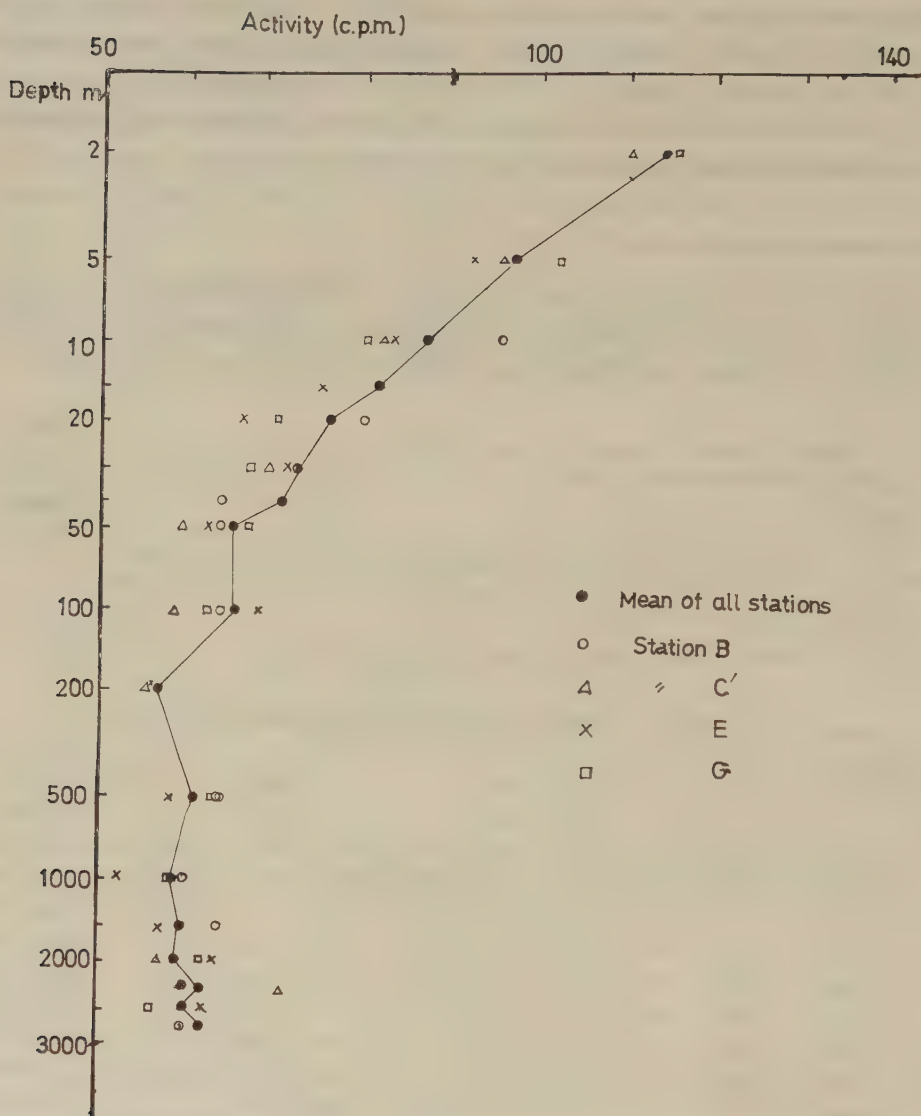


Fig. 4. Vertical distribution of γ -ray activities at some stations representative of water masses.

The fact mentioned above is probably due to the effect of cosmic ray injecting into the sea.

In the water at the depth below 200 m it could generally be said that there would be no remarkable change in activities in both directions, vertical and

horizontal. In the deep homogeneous water, the mean value of activities is as much as 60 c.p.m., the possible error of measurement being 3.5 c.p.m.. So the value larger than 65 c.p.m. is probably ascribed to the artificial contamination.

Vertical changes with depth was found in activities in deep layers particularly at station E. In the area around station E the complicated vertical construction of water masses with different origins are considered as a reason of their fluctuation.

Station C, C' and C'' are closely located in a certain area where the radioactive waste has been thrown several times by the present time. The activity in the bottom layer at station C' was found to be 71 c.p.m., which must be considered to indicate artificial contamination. The activities in the just upper layer of the bottom might not be observed at station C and C''. Further measurement must be continued in future in the area around station C for clarifying the state of artificial contamination.

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A Deep Current Measurement in the Honshu Nankai, the Sea South of Honshu, Japan*

by

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(Received October 31, 1965)

Abstract: A 4-day series of measurements of deep current with a neutrally-buoyant float at 1000 m depth was made at 28°03'.0 N, 138°02'.5 E. at a distance of 420 km of the Ogasawara Islands.

The mean of the current measurements was 2.5 cm·sec⁻¹ towards 115° true during the first 20 hr, after that it changed its velocity by some acceleration and became as fast as 7.0 cm·sec⁻¹ towards 95° true. It also exhibited a significant inertia oscillation of about 25.5 hr period during the first 50 hr superposed on the mean flow, but it did no oscillation for the next 14 hr.

Nevertheless they suggest that there are the southward deep water flows along the both sides of the Izu-Ogasawara Ridge considering the former results as previously advanced (NAN-NITI, AKAMATSU, NAKAI and FUJII, 1964; NAN-NITI, AKAMATSU and NAKAI, 1964).

1. Introduction

The former current measurements showed that a part of the upper of the south-going deep water along the Japan Trench crosses the Izu Submarine Ridge at the deep gap of its southern end (NAN-NITI, AKAMATSU, NAKAI and FUJII, 1964; NAN-NITI, AKAMATSU and NAKAI, 1964). A further observation was done in the Honshu Nankai, the anchored buoy was set at the location 28°03'.0 N, 138°02'.5 E, the depth was about 4300±50 m within a radius of 10 miles (Fig. 1). A relative error of float's position to the anchored buoy is generally less than 0.4 km (NAN-NITI *et al.*, 1964).

2. Observation

During 19th–22nd May 1965, a float was used in making the current meas-

* JEDS Contribution No. 66

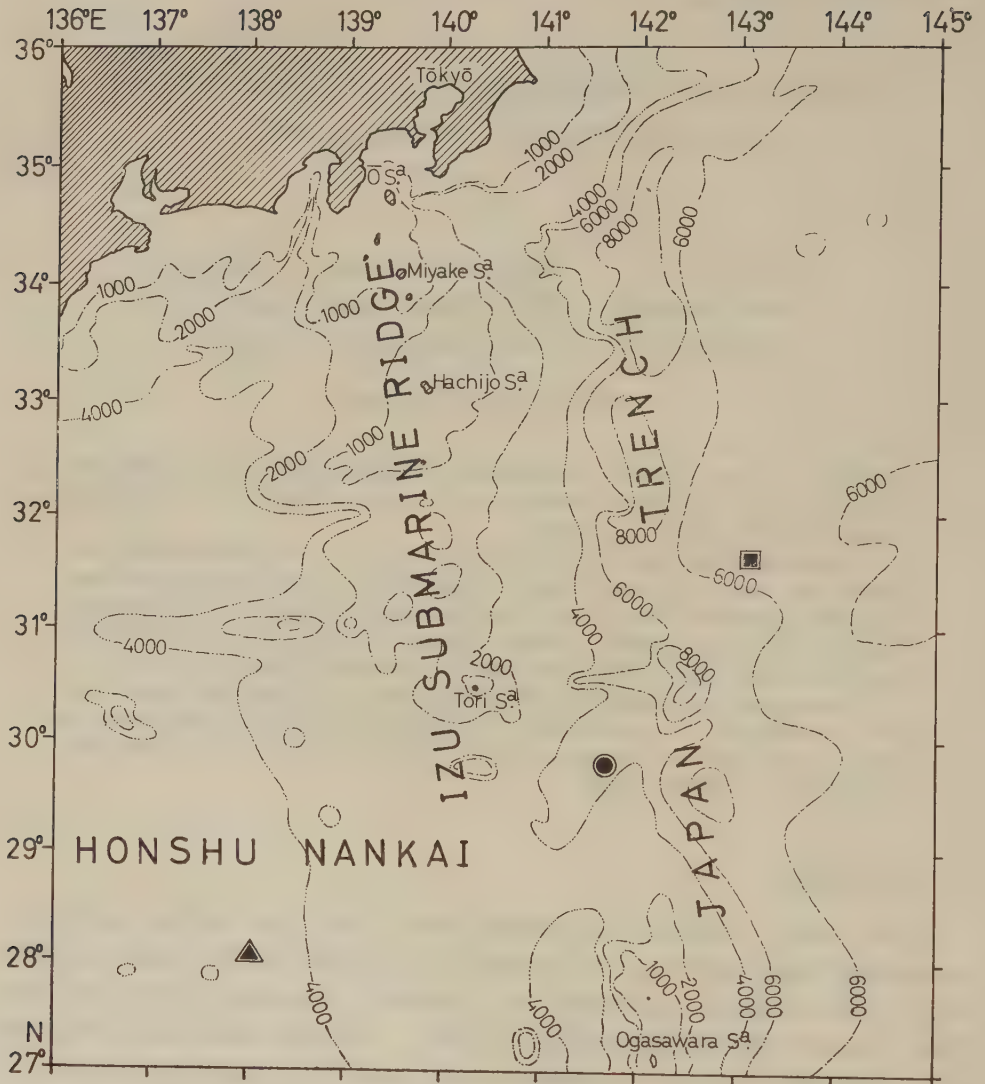


Fig. 1. Locality chart.

- ⊙ Station occupied in 1963
- Station occupied in 1964
- ▲ Station occupied in 1965

urement, loaded for 1000 m depth. The float was tracked for 64 hr, its track is shown in Fig. 2. Each dot, ○, represents a two or three-bearing fix whose circle of error is generally less than 100 m. But some dots, ×, are interpolated

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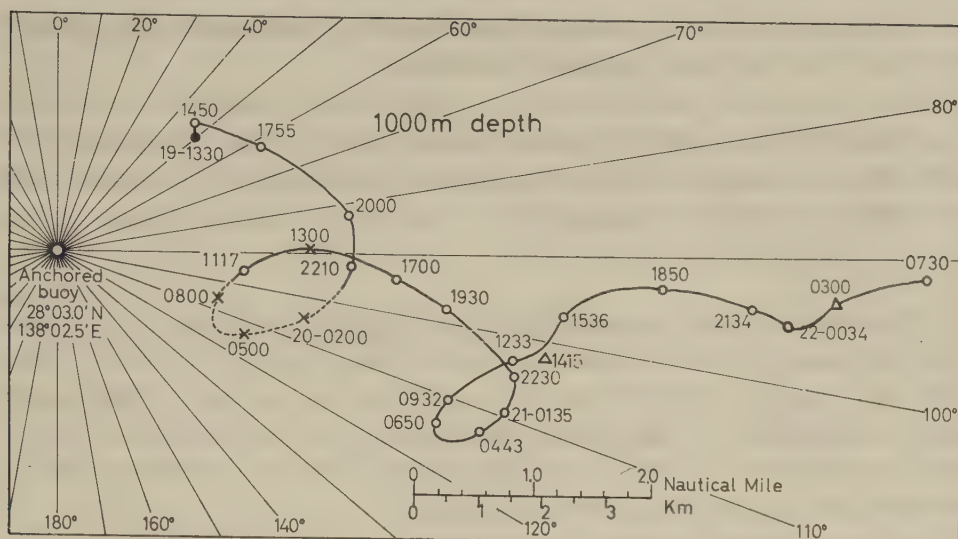


Fig. 2. Track of float. Each dot, ○, represents, a two or three-bearing fix, the date and time of each fix is indicated. A dot, △, represents only one fix, × is determined by dead reckoning method.

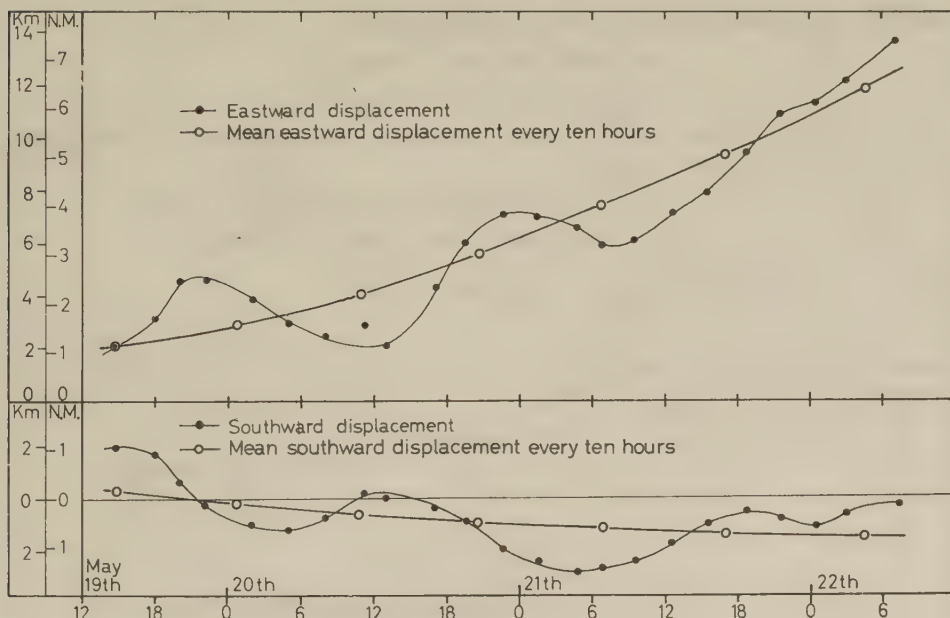


Fig. 3. Components of observed and calculated mean displacement of float plotted against time.

by dead reckoning method considering its inertia motion and two dots, Δ , represent only one fix so they may have some uncertainty.

It is noteworthy that the character of the current during the first 20-30 hr is altogether different from that during the second 45-35 hr. During the first half, there is relatively oscillative displacement; the second half is characterized by a nearly continuous straight run.

3. Results

The track shows an accelerated drift plus an oscillative motion. This means that there are an inertia motion as previously observed (NAN-NITI *et al.*, 1964) and some external force driving the water mass, so the track of float depicts the accelerated motion. The translational components of the eastward and southward displacements are obtained by least squares as follows, respectively (Fig. 3). Eastward component (taking eastward positive)

$$x = 117 \times 10^{-2} + 349 \times 10^{-4}t + 876 \times 10^{-6}t^2 \quad (\text{in mile-hour unit})$$

Southward component (taking northward positive)

$$y = 987 \times 10^{-4} - 208 \times 10^{-4}t + 936 \times 10^{-7}t^2 \quad (\text{in mile-hour unit})$$

Their velocities and accelerations are obtained as follows.

Velocity and acceleration of eastward displacement

$$u = 1.7 + 0.24 \times 10^{-4}t \quad \text{cm} \cdot \text{sec}^{-1}$$

$$a_x = 0.24 \times 10^{-4} \quad \text{cm} \cdot \text{sec}^{-2}$$

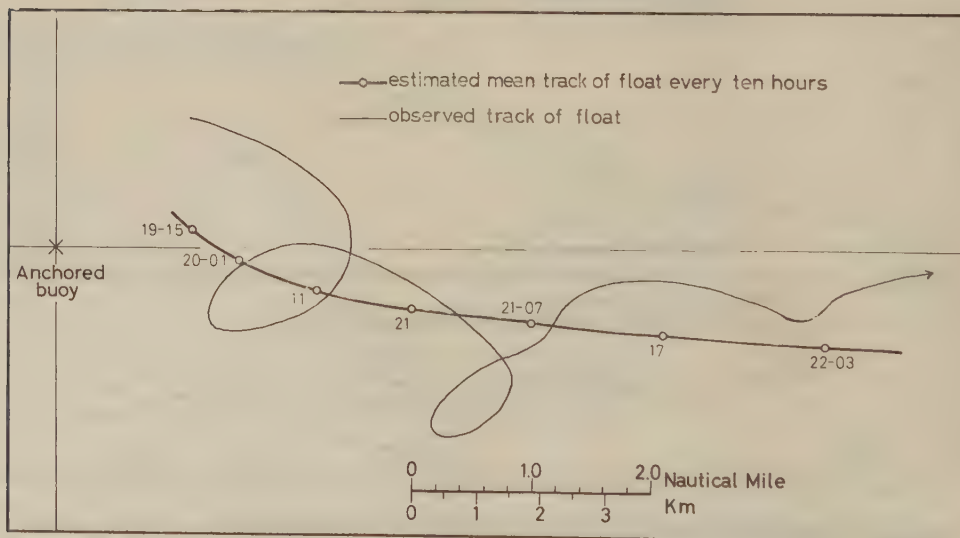


Fig. 4. Comparison of observed and estimated mean track of float.

Velocity and acceleration of southward displacement

$$v = -1.0 + 0.25 \times 10^{-5} t \quad \text{cm} \cdot \text{sec}^{-1}$$

$$a_y = 0.25 \times 10^{-5} \quad \text{cm} \cdot \text{sec}^{-2}.$$

This mean translational track is shown in Fig. 4. The mean velocity was about $2.5 \text{ cm} \cdot \text{sec}^{-1}$ towards 115° true during the first 20 hr, after that the float was accelerated by some external force and changed its velocity, it became as fast as $7.0 \text{ cm} \cdot \text{sec}^{-1}$ towards 95° true. Still it has the southward component of displacement.

After the translational component has been removed, the orbital motion is

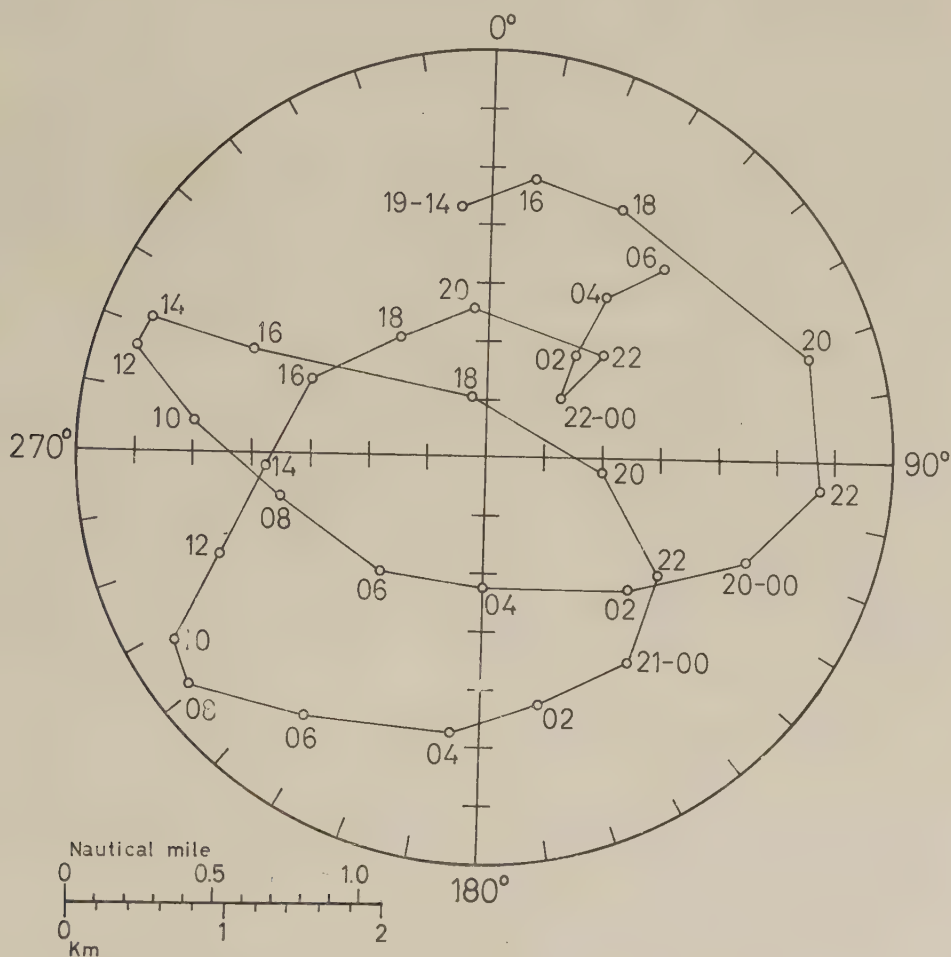


Fig. 5. Orbital motion of float. The unit of co-ordinates is nautical mile.

shown in Fig. 5. It moved in a clockwise elliptical orbit at speed of $8\text{--}20\text{ cm}\cdot\text{sec}^{-1}$, at the first cycle, with a 4.2 km of the major axis and 2.6 km of the minor one. By the effect of some external force, the next cycle showed a rather smaller and more circular orbit with a 2.6 km of the major axis and a 2.2 km of the minor one. Their periodicity seems to be about 26 hr. The theoretical value of half pendulum day at $28^{\circ}03'$ latitude, the mean location of observation, is 25.5 hr, and its diameter of the circular motion is 2.4–5.8 km provided its speed to be $8\text{--}20\text{ cm}\cdot\text{sec}^{-1}$. The calculated diameter is a little larger than the observed ones.

The inertia circular motion was damped by some external force; the force, for instance, might be revealed as the accelerating translational force. The time change of the mean diameter of the circle can be written as

$$D = D_0 e^{-\alpha t} = 3.4 \times e^{-1.3 \times 10^{-2} t} \quad (\text{in km-hour unit}).$$

So the expected diameter at 19h on 21st, it is the time which 51 hr passed after 16h on 19th, will be 1.7 km. But such a circulative motion could not be observed. Though failing to finding the float till 07h30 m after 00h34 m fixing, still the position of float at 07h30 m showed that there is no such a track as the inertia oscillation will depict considering its speed and periodicity (Fig. 2).

4. *Estimating of the depths of floats*

Two or three bearings were taken from different ship's positions to decide a float's position in horizontal plane assuming the constant depth of the float to be 1000 m. Conversely, the scattering of bearings could estimate the depth of float to be about 1000 ± 200 m assuming the some definite distance in horizontal plane between the ship and the float.

By another method, the expected depth which the float was made neutrally-buoyant by added weights was calculated to be 1006 m from observations of temperature and salinity.

5. *The method of computing the velocity profiles*

At five hydrographic stations, $I_{22}\text{--}I_{26}$, temperatures, chlorinities and quantities of dissolved oxygen were observed (Figs. 6, 7). The dynamically calculated velocities of sections A, B, C referred to 1200 m depth, the oxygen minimum layer, are shown in Fig. 6.

The float observation showed the eastward flow at 1000 m depth with a speed of $2\text{--}7\text{ cm}\cdot\text{sec}^{-1}$, and the photographic bottom current measurement (SASAKI,

T. Nan-niti and others: A Deep Current Measurement in the Honshu Nankai

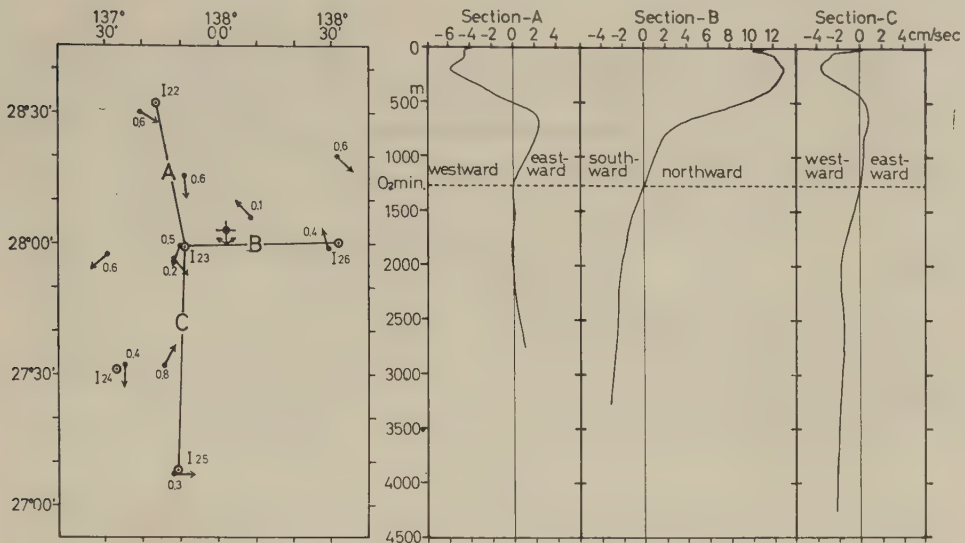


Fig. 6. Hydrographic stations surrounding the anchored buoy, and geostrophic currents in each section.

WATANABE, OSHIBA and KISHINO, 1965) showed the northward current at 4250 m depth, 331° true, with a speed of $2.5 \text{ cm} \cdot \text{sec}^{-1}$. By the profiles of section A and C, we can see the eastward current at 1000 m depth, though the both profiles are different from each other; but we can hardly expect the observed bottom current by the dynamic calculations.

The surface current with the GEK observations is very complicated (Fig. 6). The each current was not observed at the same time but it took about one week to complete the series of observations. Now we have Fig. 8 which shows the time variation of the surface and the deep water current; in the figure the deep water current shows only its direction, and the surface current shows its speed and direction.

From the figure, we have a conclusion that the surface current observed by the GEK shows the periodic change as like as the deep current. Of course it will be natural that the surface current just above the float will have the same phase with the deep current. The figures (Figs. 6, 8) show that the whole area of about 90 miles square had the same phase. Was this periodic current caused by the tidal motion? No semi-diurnal period was observed, so this phenomenon seems to be caused by the inertia motion, as previously concluded (NAN-NITI, AKAMATSU, NAKAI and FUJII, 1964).

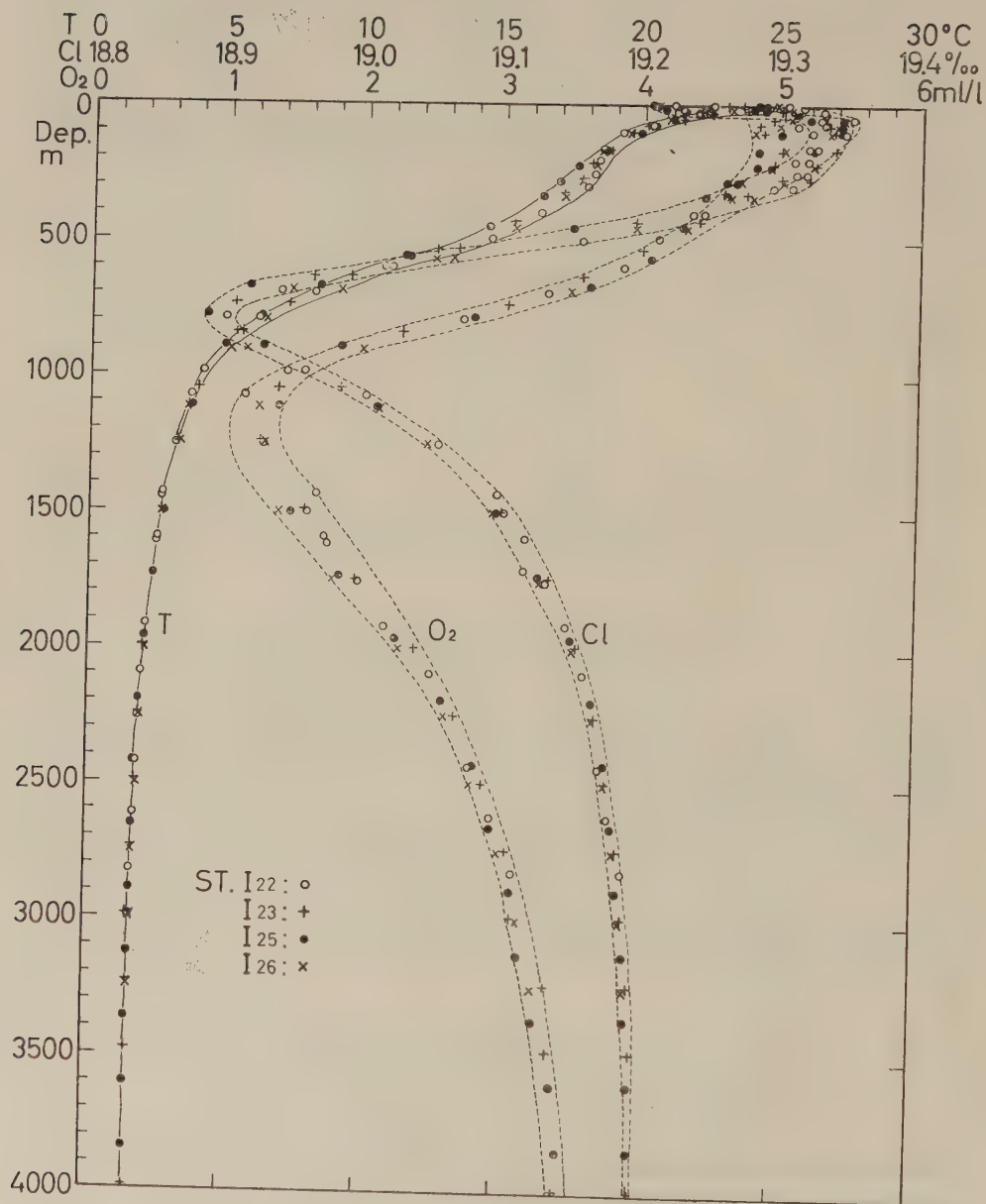
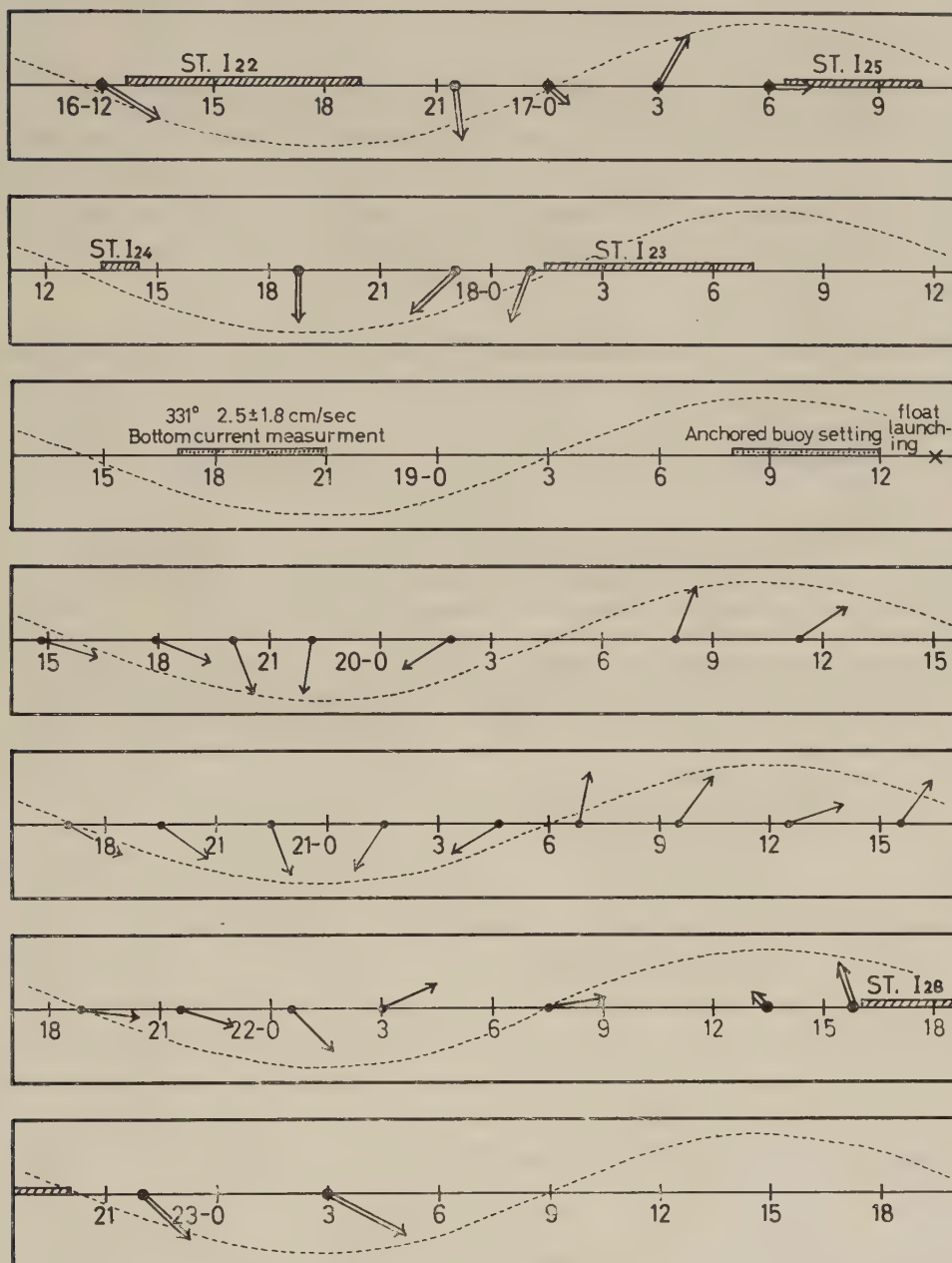


Fig. 7. Vertical distributions of temperature, chlorinity, dissolved oxygen.



- surface current
- deep water current

Fi. 8. Time variation of the surface and the deep water current.

6. Conclusion

With the results obtained by this and the former observations (NAN-NITI *et al.*, 1964), we have a following conclusion that there are the south going components of the deep current on the both sides along the Izu-Ogasawara Submarine Ridge. These observations exhibited significant inertia oscillations, and in this time after the period of two cycles, the oscillation disappeared from the observation by some external force.

Acknowledgements: The authors wish to express their sincere appreciation for the cooperation of member of observation and the crew on board the R.S. Ryofu-Maru.

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Numerical Computations for the Chilean Earthquake Tsunami

By

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(Received November 2, 1965)

Abstract: Numerical computations are made for the Chilean Earthquake Tsunami, using the five degree mesh points and the real depth of all the Pacific Ocean. Being the small capacity and the slow speed, our computer IBM 704 is not enough to study the convergence or divergence of tsunami energy. But the obtained results show that the arrival time is comparatively in good agreement with the refraction diagram and why the amplitudes near Japan, Hawaii Is. and Alaska are big are shown in connection with the tidal theory.

Tidal waves travelling over the rotating globe were numerically computed by UENO (1964 a, b, 1965) using the grid points of 5 degree interval. The results tell us this method is very useful to study large scale phenomena like tidal waves. That is, in the formulas (18)~(22) of UENO's paper (1964 a) if we put $Hm=0$ or $D=0$, they reduce to the prediction equations for free gravity waves travelling over the rotating spherical surface. One of good examples of this type may be the Tsunami, especially, of the Chilean Earthquake occurred on May 22, 1960, which has the origin in the southern Chile and hit all the Pacific coast, having the 5.8 m highest waves on the northern Japanese coast (J.M.A., 1961).

The techniques of the computations are very simple as follows: In Fig. 3 (b) of UENO's paper (1964 a) both western and eastern boundaries are closed. At the initial state the elevations and currents at all the grid points are put to be zero, except, at the seven points along the coast of Chile which represent the origin (Fig. 1). The initial elevation at the origin must satisfy the relation $E = \frac{1}{2} \rho g \zeta_0^2 S$, where ρ is the density of sea water, g is the gravity and E is the total energy of the tsunami estimated at $(1\sim 3) \times 10^{23}$ ergs from the tidal records of Christmas and Johnston Is. (J.M.A. 1961), S is the area of origin represented by the above seven grid points, that is, as each elevation grid point should occupy the area $2 \times$ (five degree square), $S = 3 \times 10^{16} \text{cm}^2$. Then ζ_0 becomes about 120 cm. Moreover,

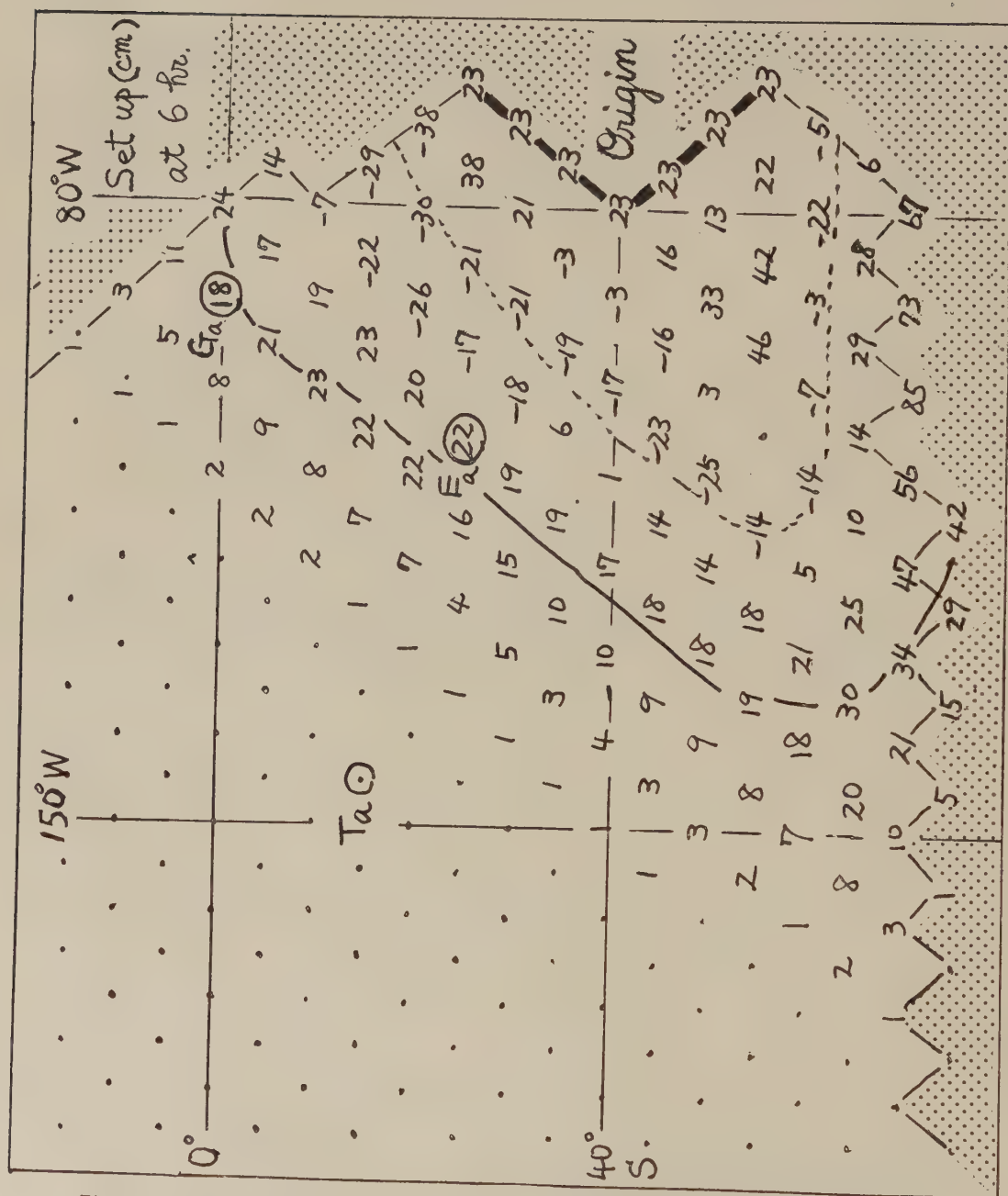


Fig. 1. Elevation distribution after 6 hrs, full line is crest and dotted line trough of first wave. Ea is Easter, Ga, Galapagos and Ta, Tahiti Is.

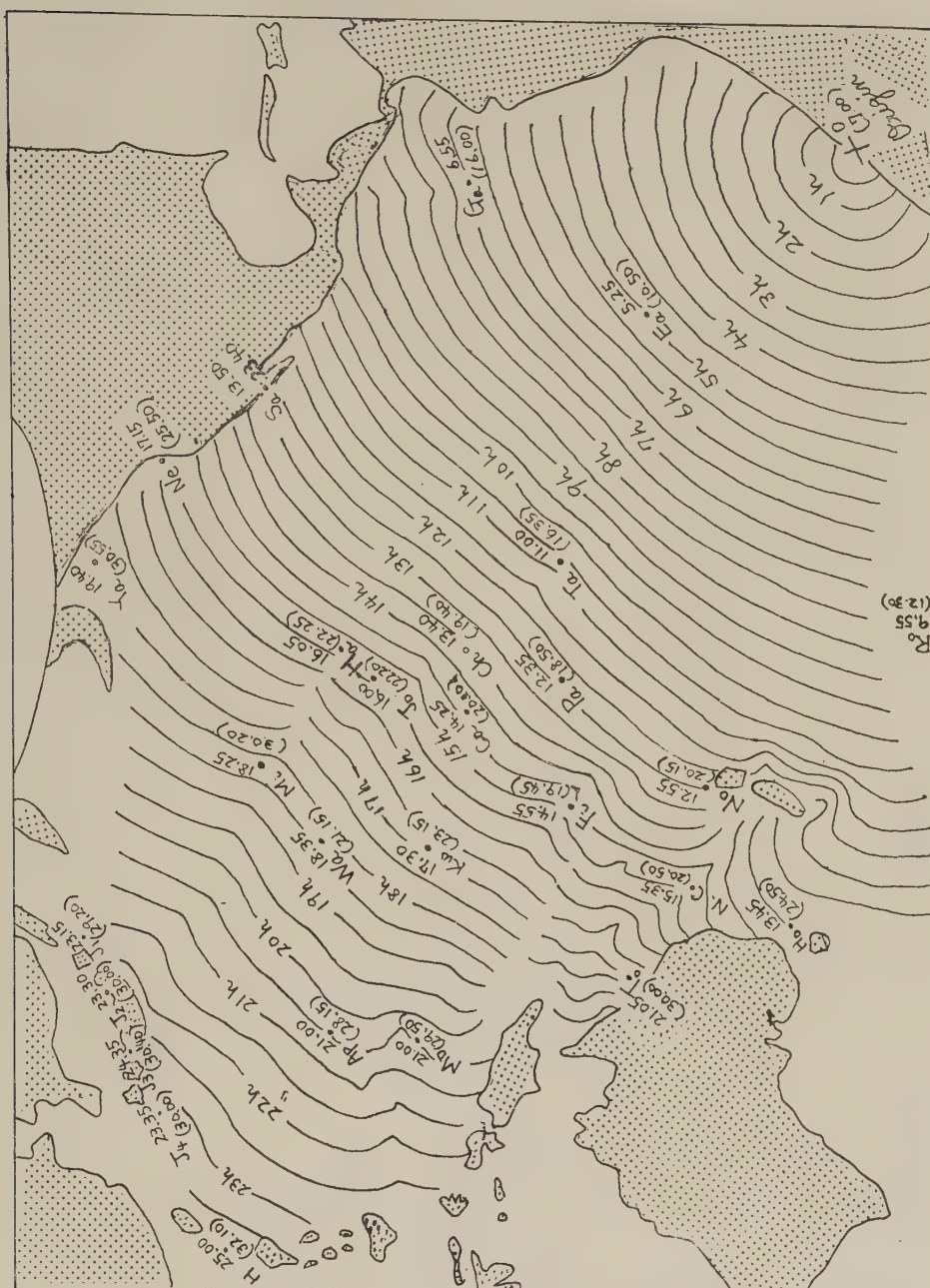


Fig. 2. Arrival time. Figures are computed first and second wave (in parentheses). Curves are from refraction diagram. (Station names are shown in Table 1.)

in connection with the conjugate lattices (UENO, 1964c, §5.6), the following computational technique is done: at every time step, along seven grid points on the origin the influences from the origin should be equal as shown in Fig. 1.

The brief explanations for computed results are done by Fig. 1-5. with the station names in the Table 1. Fig. 1 shows that after 6 hr the first wave reaches

Table 1. Station name

Abb.	Name	Abb.	Name
Ga	Galapagos	Ro	Ross Sea
Sa	San Francisco	No	Nofolk
Ne	Neah Bay	Ho	Hobert
Ya	Yakutat	N. C.	New Castle
Ea	Easter	Fi	Fiji
Ta	Tahiti	To	Townsville
Pa	Pago	Mo	Moen
Ch	Christmas	Ap	Apra
Ca	Canton	J ₁	Japan
Ha	Hawaii	J ₂	Japan
Jo	Johnston	J ₃	Japan
Kw	Kwajalein	J ₄	Japan
Wa	Wake	H	Hong Kong
Mi	Midway		

between Easter and Galapagos Is., the second wave at the origin and that the influences of reflection from the Antarctic appear near the southern boundary. We can easily know from the elevation near the Antarctic in Fig. 1 that the zigzag boundary causes the set up distribution between conjugate lattices so rugged features that the reflections from both north and south boundaries, as expected from Fig. 3 (b) of UENO's paper (1964a) make too difficult to analyse the loop or node of seiche of the Pacific. Fig. 2 is the arrival times, the curves are the estimated ones from the refraction diagram (J.M.A., 1961) and the figure at each island means the computed arrival time of first wave and the second (in parentheses). The discord between the two is chiefly caused by the fact that the computational locations of origin and islands are measured by 5 degrees. To pass through 5 degree interval the long wave travelling over the 4000 m deep ocean takes 45 minutes.

Fig. 3 is the distribution of the first and second (in parentheses) wave set up (in cm.) from the mean sea level. These absolute values have no meaning,

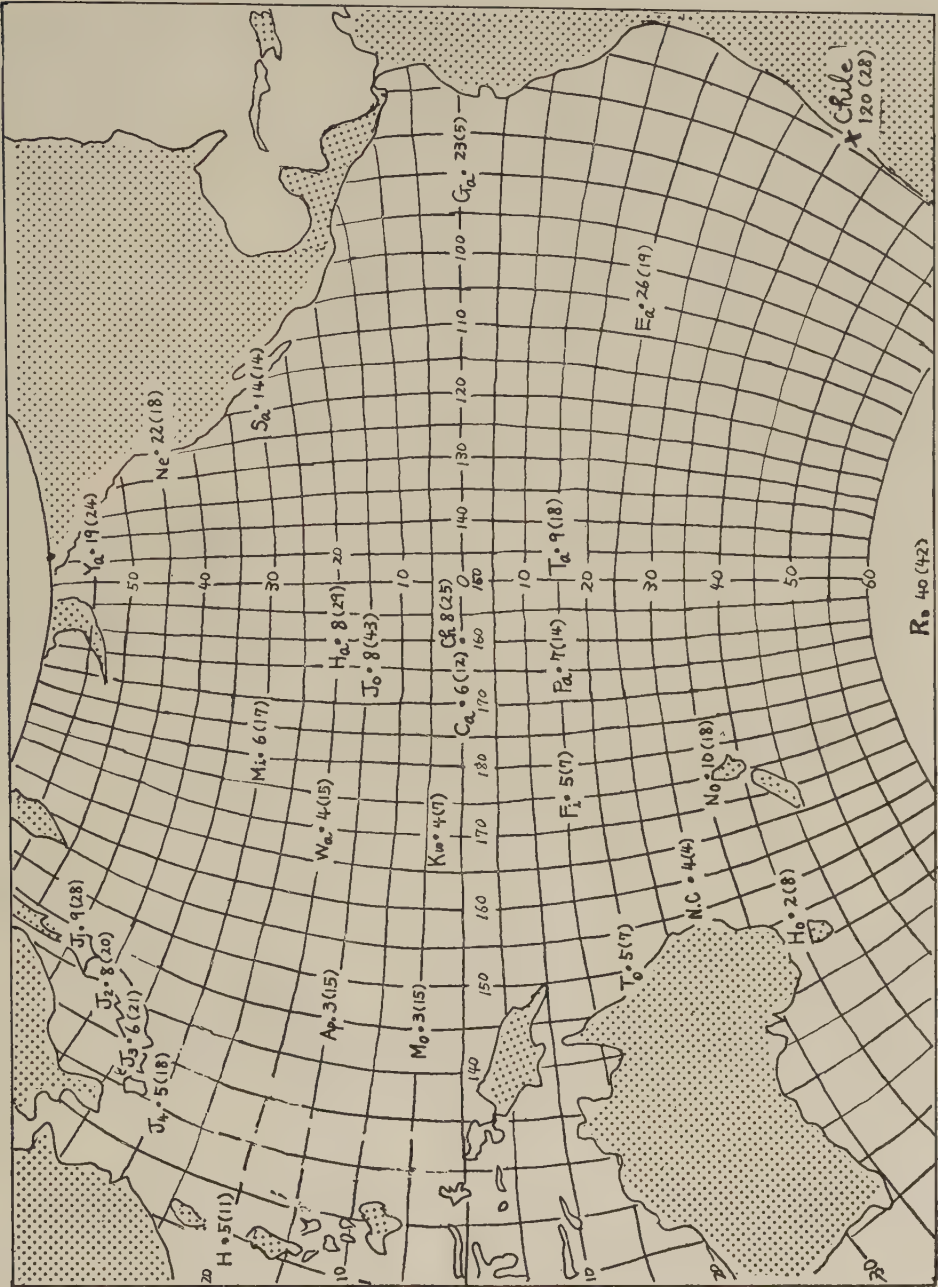


Fig. 3. First set up and second one (in parentheses). (Station name are shown in Table 1.)

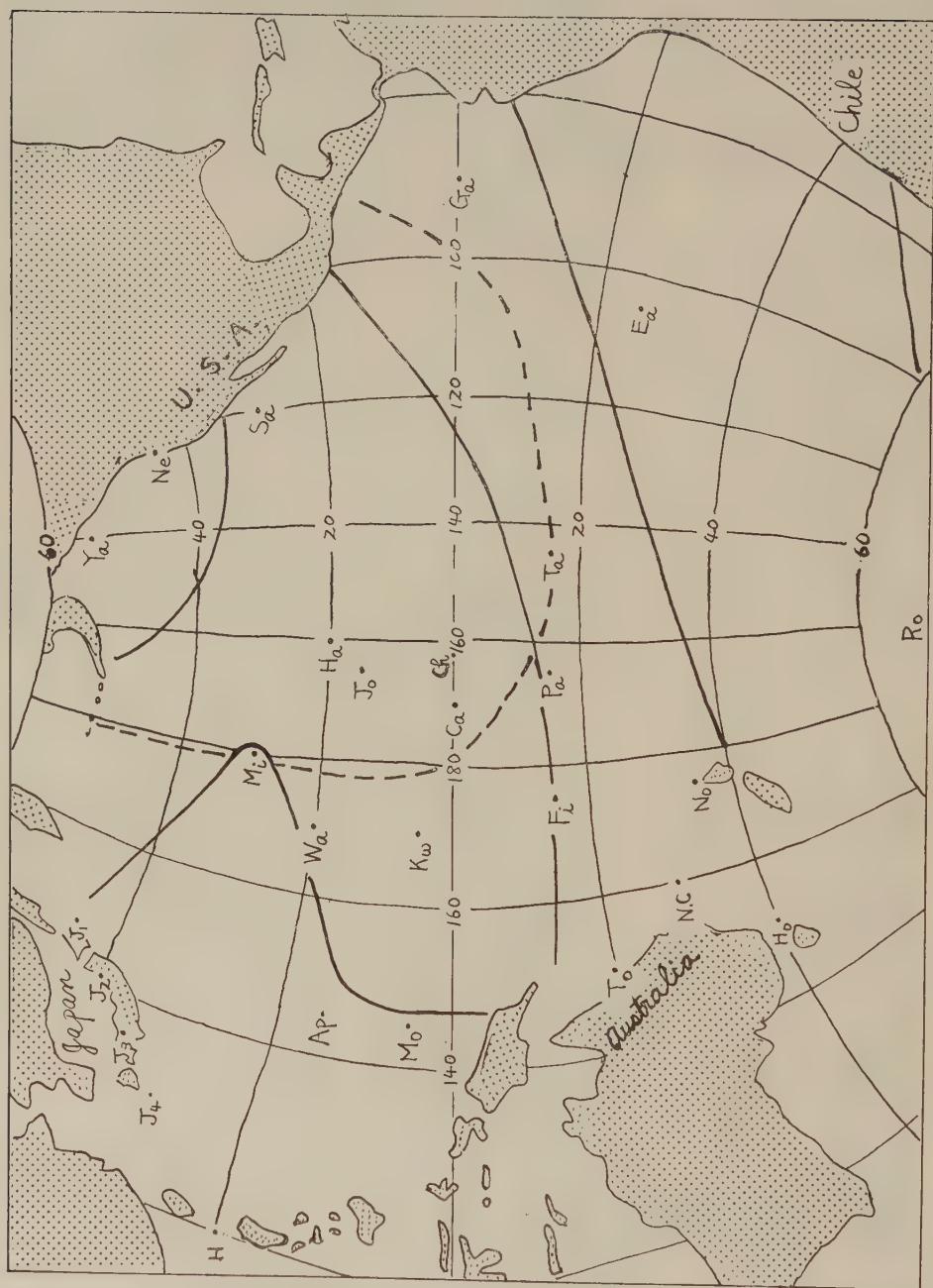


Fig. 4. Tidal seiche. Dotted line is nodal line for diurnal tide (K_1, O_1) and full line for semi-diurnal tide (M_2). (Station names are shown in Table 1.)

because the energy dissipation and the effects of continental shelves are not included in the computations. If we regard them as the relative values, however, we can notice very interesting two facts. One is that the first set up decreases according to the distance from the origin, however, again increases on the coast of Japan and Alaska. Another is that the large second set up is computed along the coast of Japan, Alaska and near the Hawaii Is. and the small one near Midway and Wake Is.. This fact reminds us of the seiches of the Pacific, and to analyse the phenomena by our computations is very difficult as mentioned above. But in the tidal theory this is well studied by both observation (BOGRDANOV, 1962) and numerical studies (UENO, 1964a, b, 1965). According to their results, the author thinks that the Pacific Ocean can be regarded as a rectangular basin of 11,000 Km width (from Australia to U.S.A.) and of 17,000 Km length (from Japan to Chile) (Fig. 4). As drawn in Fig. 4, we can see the quite definite lateral seiche of 24 hr period, which is corresponding to K_1 or O_1 tide. On the other hand the longitudinal seiche is not so clear as the lateral one, especially in the southern part, and the Pacific Ocean is divided into several part so that each part has a 12 hr period corresponding to M_2 or S_2 tide. In Fig. 4 the dotted line shows the nodal line of K_1 or O_1 and the full line of the M_2 (O_1 's nodal line is slightly different from Fig. 4). According to the Fig. 4, the common loops of diurnal and semi-diurnal tide exist adjacent to Japan, Alaska, Hawaii, Johnston and Christmas Is., and if the Chilean Earthquake causes the seiche in the Pacific Ocean, we can see the big amplitude in the areas as mentioned above. Fig. 5 shows the curve of computed elevation at each station, where the 6 hr period is predominant. At Wake Is. the amplitude is small and at Japan and Hawaii Is. they are big. Further the phase between Japan and Hawaii Is. is opposite each other as expected from Fig. 4.

The results obtained by this computation are very insufficient to analyse the convergence or divergence of tsunami's energy. If we want to study this important problem, we must use more large computer enough to consider the continental shelves where most of the energy loss occurs or to represent the topography of the continental coasts and the origin more correctly.

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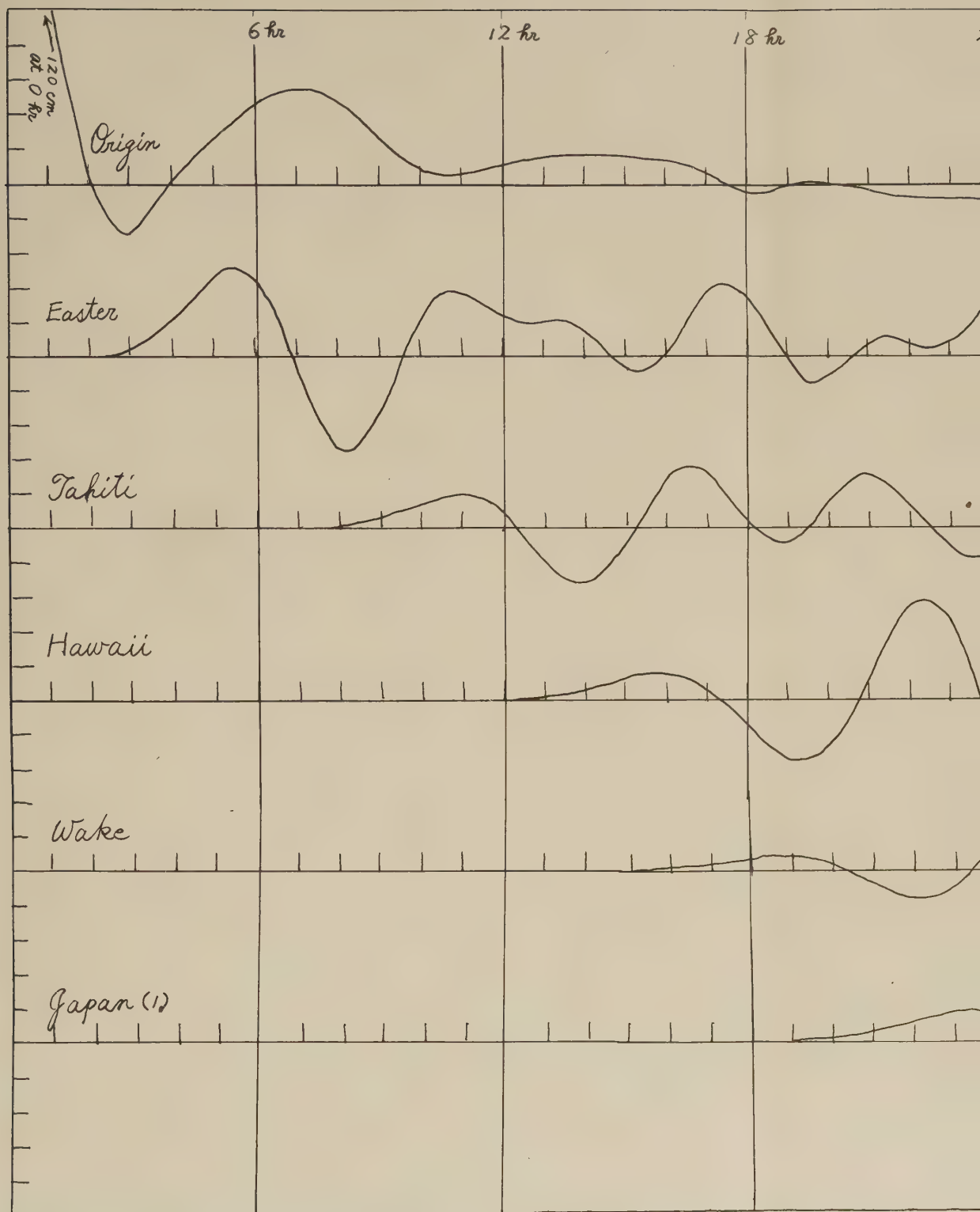
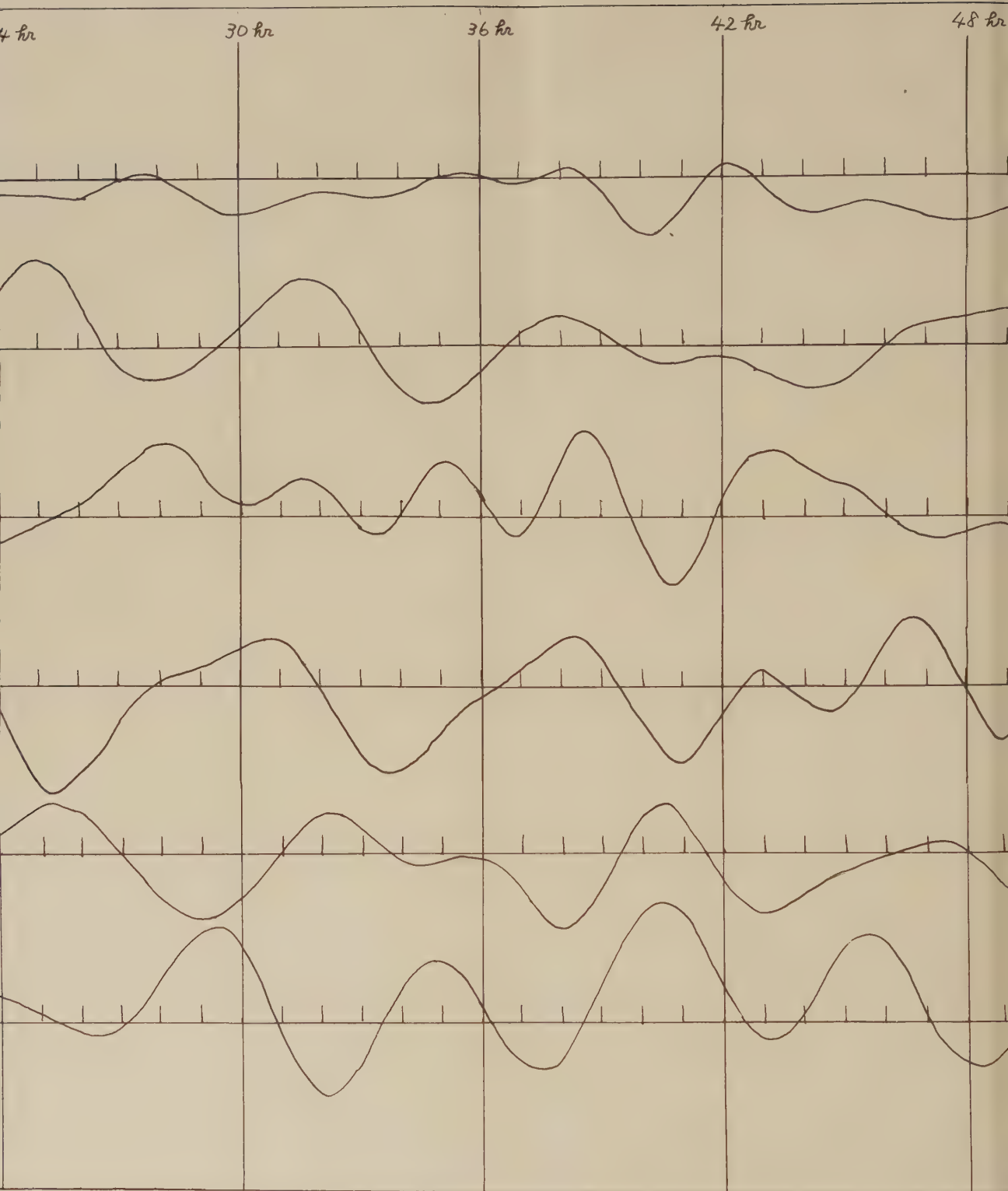
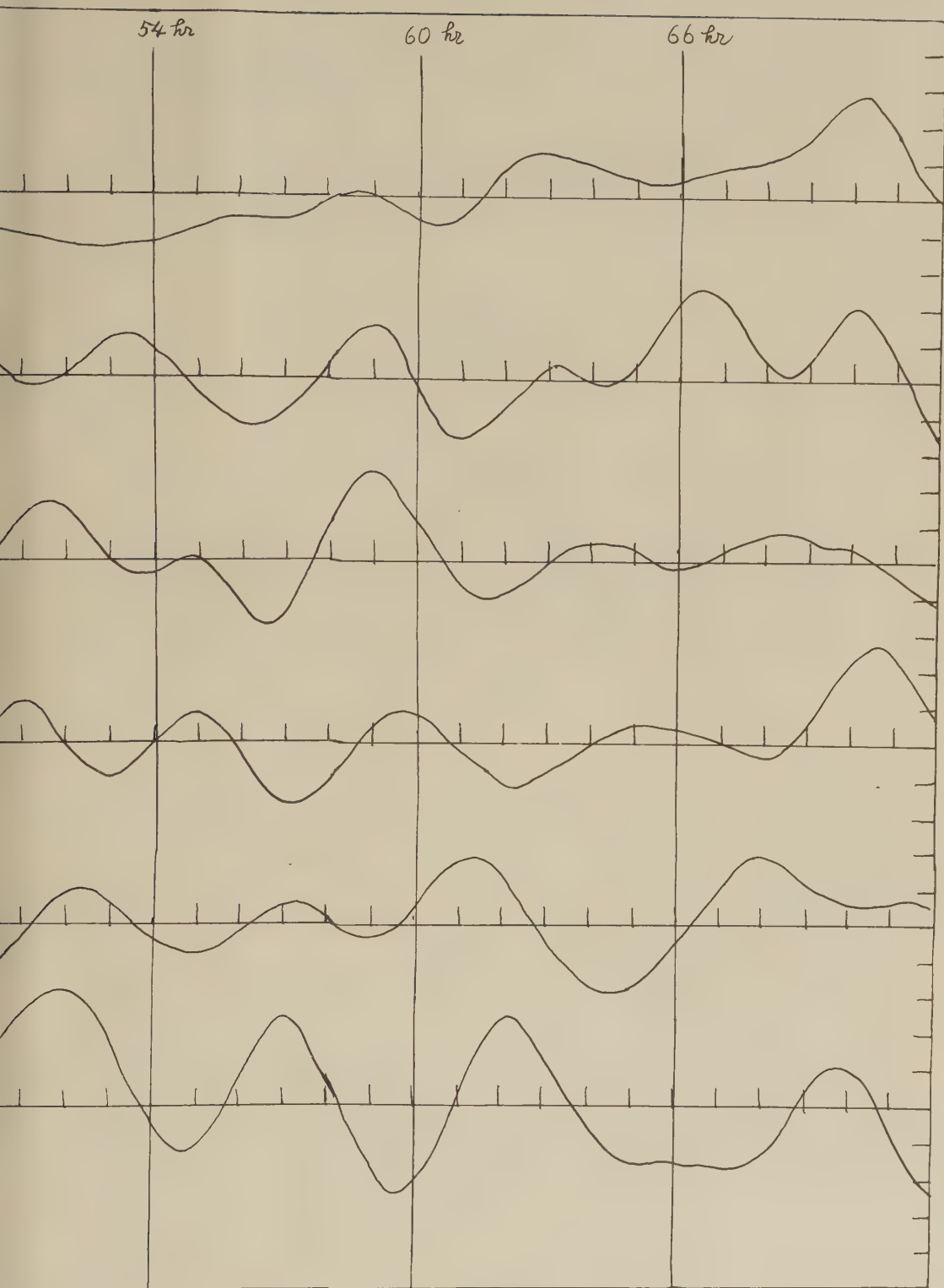


Fig. 5. Variatio



of set up at several stations, abscissa shows the time in one hour unit and ordinate the set up in



unit.

The Distribution of the Four Boreal Species of Copepoda in the Western North Pacific*

By

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(Received November 2, 1965)

Abstract: It has been well known that such species of copepods as *Calanus cristatus*, *C. plumchrus*, *Eucalanus bungii bungii* and *Metridia lucens* were abundant in the Oyashio cold water. The distribution of them, however, has not been clarified obviously in the sea area south of the Kuroshio Extension. In the present survey, it has been cleared up that these species were dispersed far southwards from the Kuroshio Extension, especially *Eucalanus bungii bungii* reached to the south-west of the Ogasawara Submarine Ridge, though the number of individuals was extremely scarce. The layer which the copepods were distributed mainly was from 500 to 1000 m depth. This fact indicates that there is a southward flow originating in the Oyashio water in relatively deeper layer.

1. Introduction

Referring to the data published by the Japan Meteorological Agency (1955-1962), such four copepods as *Calanus cristatus*, *C. plumchrus*, *Eucalanus bungii bungii* and *Metridia lucens* are found to be the main constituents of the large copepod community in the Oyashio region. It is well known that the bio-geographic southern boundary of the distribution of boreal copepods is the Kuroshio Extension flowing eastwards in the east of Japan. The knowledge, however, of their distribution in the deep water off southern Japan is extremely poor, because the deep-sea plankton collections have been scarcely operated in the said waters.

Since 1959, the Japanese Expeditions of Deep Sea have been carried out on board the R. V. Ryōfū-Maru of the Japan Meteorological Agency. In these expeditions, we have collected many deep-sea plankton samples. In addition to those samples, other samples were taken in the cruises of the R. V. Ryōfū-Maru. On the basis of these materials, the distribution of the four boreal copepods in deep water have been presented in this report.

* JEDS Contribution No. 68.

2. Methods of sampling

The number of the sampling station is twelve in all, and the location is shown in Fig. 1. The Sts. A 3, G 207, E 2, F 23, I 2, I 4 and I 23 are those of plankton

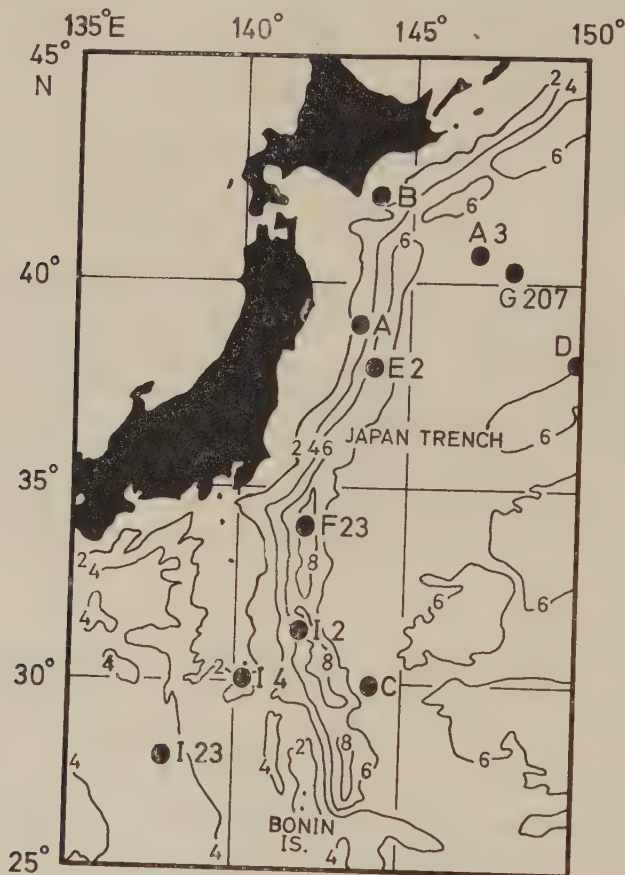


Fig. 1. Approximate location of the sampling positions.

collections made in the Japanese Expeditions of Deep Sea. At the former three stations, a 130 cm conical net (SUYEHIRO *et al.*, 1960), open type, was used making vertical hauls from deeper layers. The lower limits of hauling were 2500 m, 3000 m and 3000 m depths, respectively. It has a mouth ring of 130 cm diameter and a filtering part made of rough grit gauze with 3 mm apertures and fine bolting silk with 0.33 mm apertures.

At the latter four stations, a 130 cm closing net (MATSUE *et al.*, 1963) was

used making sectional hauls in the deep. The lower limits of the net were 6500 m, 3000 m, 1800 m and 3000 m depths, respectively. The net used was a type re-constructed from the original 130 cm conical net so as to close the mouth.

At the Sts. A, B and C, the automatic opening and closing multiple net was used, of which closing was made by the revolution of a propeller. It has a square mouth frame of 40×40 cm and a filtering part made of bolting silk with 0.3 mm apertures. The lower limit of the net was 1000 m depth in every station.

At the St. D, a hinged double vertical closing net (MOTODA *et al.*, 1962) was used, which had a square mouth frame of 40×40 cm and a filtering part made as same as it of the multiple net. In using it, continuous two sectinol collections were made vertically by one hauling. The lower limit of hauling was 750 m depth.

3. Hydrography and results of observations

The vertical distributions of temperature and chlorinity are shown in Figs. 2-4. The Sts. G 207, A 3, A and B are located in the Oyashio region, while the Sts. F 23, I 2, I 4, I 23 and C in the kuroshio region and the Sts. E 2 and D in the transitional region. Therefore, the upper water from the sea surface to about 1000 m depth showed conspicuous different temperature and chlorinity. The core of the

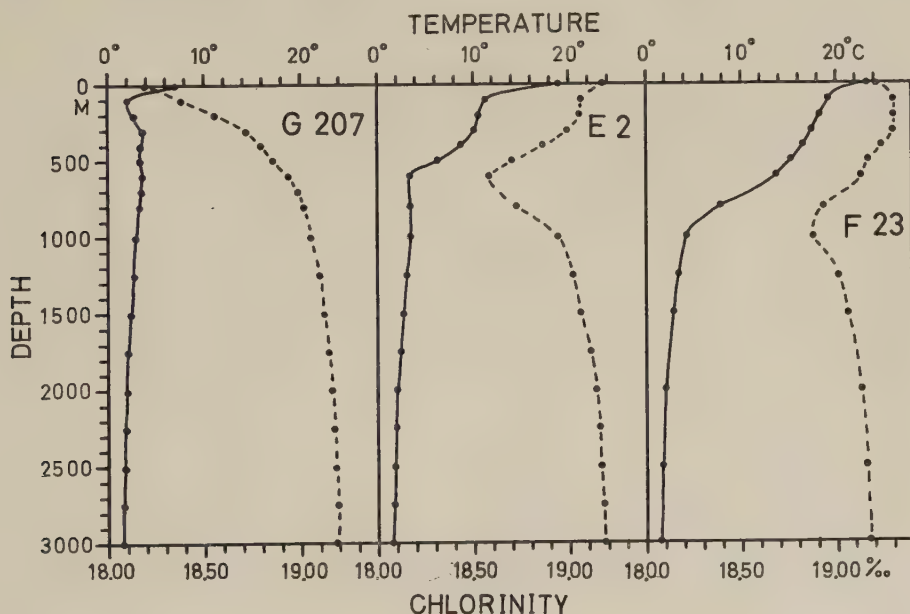


Fig. 2. Vertical distribution of temperature and chlorinity at the Sts. G207, E 2 and F 23.

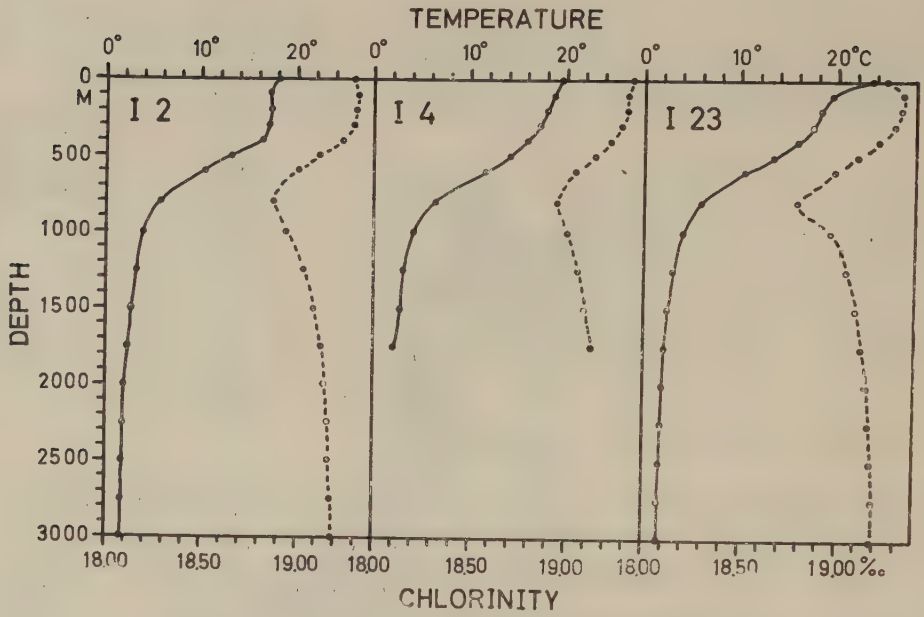


Fig. 3. Vertical distribution of temperature and chlorinity at the Sts. I 2, I 4 and I 23.

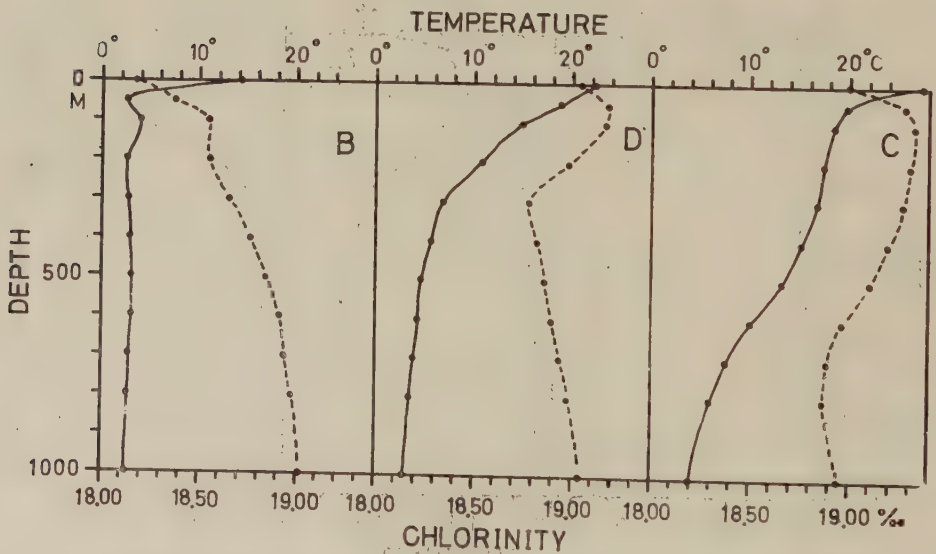


Fig. 4. Vertical distribution of temperature and chlorinity at the Sts. B, D and C.

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Subarctic Intermediate Water is shown with a chlorinity-minimum layer at the stations located in both of the Kuroshio and transitional regions.

In this paper, the numbers of individuals of various copepods are shown concerning later copepodid stages and adult, as earlier copepodid stages were not found in the southern part of the present sea area. These are shown in Tables 1-5.

Table 1. Number of individuals (per a haul) of four boreal copepods at the Sts.
A 3, G 207, E 2 and F 23.

Station	A 3	G 207	E 2	F 23
Hauling depth (m)	0-2500	0-3000	0-3000	0-5000
Wire angle	29°		15°	
Location	40°38' N 148°19' E ↓ 40°36' N 147°19' E	40°23' N 148°26' E ↓ 40°24' N 148°28' E	37°52' N 143°50' E ↓ 37°51' N 143°48' E	34°30' N 142°20' E ↓ 34°32' N 142°22' E
Date	Sep. 16th, 1959	June 17th, 1963	June 13th, 1961	June 17th, 1962
Time	11h18m ↓ 12h27m	00h59m ↓ 02h23m	16h06m ↓ 17h30m	17h59m ↓ 20h30m
<i>Calanus cristatus</i> VI V•IV	464 612	312 644	304 504	4 11
<i>C. plumchrus</i> VI V•IV	1264 1664	166 604	— 16	1 8
<i>Eucalanus bungii</i> VI <i>bungii</i> V•IV	1120 1800	270 220	472 28	60 16
<i>Metridia lucens</i> VI	572	280	56	8

The roman numerals denote the stage of copepodids.

Table 2. Number of individuals (per a haul) of four boreal copepods obtained by sectional haul at the St. F 23.

Station		F 23		
Hauling depth (m)		0-500	500-1000	1000-2000
Wire angle				
Location		34°28' N 124°17' E	34°27' N 142°16' E ↓ 34°28' N 142°17' E	34°24' N 142°15' E ↓ 34°27' N 142°16' E
Date		June 17th, 1962		
Time		10h58m ↓ 11h12m	09h47m ↓ 10h16m	06h44m ↓ 08h40m
<i>Calanus cristatus</i>	VI V•IV	— —	3 3	3 35
<i>C. plumchrus</i>	VI V•VI	— —	— —	— —
<i>Eucalanus bungii</i> <i>bungii</i>	VI V•VI	— —	1 —	24 —
<i>Metridia lucens</i>	VI	—	13	—

Station		F 23		
Hauling depth (m)		2000-3000	3000-4000	5000-6500
Wire angle		47°		
Location		33°59' N 142°03' E ↓ 34°07' N 142°12' E	34°29' N 142°17' E ↓ 34°30' N 142°20' E	34°32' N 142°22' E ↓ 34°35' N 142°22' E
Date		June 15th, 1962	June 17th, 1962	June 18th, 1962
Time		19h02m ↓ 20h30m	12h17m ↓ 15h15m	00h18m ↓ 03h58m
<i>Calanus cristatus</i>	VI V•IV	1 9	— —	— —
<i>C. plumchrus</i>	VI V•IV	— —	— —	— —
<i>Eucalanus bungii</i> <i>bungii</i>	VI V•IV	7 —	— —	— —
<i>Metridia lucens</i>	VI	—	—	—

The roman numerals denote the stage of copepodids.

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Table 3. Number of individuals (per a haul) of four boreal copepods obtained by sectional haul at the Sts. I 2, I 4 and I 23.

Station	I 2			
Hauling depth (m)	0-500	500-1000	1000-2000	2000-3000
Wire angle	30°			
Location	31°21' N 142°00' E	31°22' N 141°58' E	31°24' N 141°56' E	31°26' N 141°53' E
Date	April 26th, 1965			
Time	11h46m ↓ 12h27m	13h10m ↓ 14h12m	14h44m ↓ 16h35m	16h55m ↓ 19h25m
<i>Calanus cristatus</i> VI V•IV	— —	— —	— —	— —
<i>C. plumchrus</i> VI V•IV	— —	— —	— —	— —
<i>Eucalanus bungii</i> VI <i>bungii</i> V•IV	— —	1 —	— —	— —
<i>Metridia lucens</i> VI	—	—	—	—
Station	I 4			
Hauling depth (m)	0-500	500-1000	1000-1800	
Wire angle	0°			
Location	30°54' N 140°14' E	30°54' N 140°14' E	30°54' N 140°15' E	
Date	April 28-29th, 1965			
Time	20h24m ↓ 21h10m	21h34m ↓ 22h28m	22h45m ↓ 00h14m	
<i>Calanus cristatus</i> VI V•IV	— —	— —	— —	
<i>C. plumchrus</i> VI V•IV	— —	1 —	— —	
<i>Eucalanus bungii</i> VI <i>bungii</i> V•IV	— —	— 1	1 —	
<i>Metridia lucens</i> VI	—	—	—	

Station	I 23			
Hauling depth (m)	0-500	500-1000	1000-2000	2000-3000
Wire angle	0°			
Location	27°58'N 137°52'E	27°59'N 137°52'E	28°01'N 137°52'E	28°01'N 137°52'E
Date	May 18th, 1965			
Time	08 ^h 40 ^m ↓ 09 ^h 10 ^m	09 ^h 28 ^m ↓ 10 ^h 30 ^m	10 ^h 45 ^m ↓ 12 ^h 30 ^m	12 ^h 45 ^m ↓ 15 ^h 18 ^m
<i>Calanus cristatus</i> VI V•IV	— —	— —	— —	— —
<i>C. plumchrus</i> VI V•IV	— —	— —	— —	— —
<i>Eucalanus bungii</i> VI bungii V•IV	— —	2 1	— —	— —
<i>Metridia lucens</i> VI	—	—	—	—

The roman numerals denote the stage of copepodids.

Table 4. Number of individuals (per a haul) of four boreal copepods obtained by sectional haul at the Sts. A, B and C.

Station	A		
Hauling depth (m)	0-500	500-750	750-1000
Wire angle	45°		
Location	39°-04'N 143°-35'E		
Date	June 30th, 1965		
Time	13 ^h 05 ^m — 13 ^h 57 ^m		
<i>Calanus cristatus</i> VI V•IV	— 153	— 16	12 8
<i>C. plumchrus</i> VI V•IV	4 3715	208 288	144 424
<i>Eucalanus bungii</i> VI bungii V•IV	1 320	168 528	12 138
<i>Metridia lucens</i> VI	588	992	184

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Station	B		
Hauling depth (m)	0-500	500-750	750-1000
Wire angle	10°		
Location	41°-58' N 143°-58' E		
Date	July 8th, 1965		
Time	11 ^h 45 ^m — 12 ^h 45 ^m		
<i>Calanus cristatus</i> VI V•IV	— 64	2 5	12 56
<i>C. plumchrus</i> VI V•IV	— 701	56 200	80 132
<i>Eucalanus bungii</i> VI <i>bungii</i> V•IV	— 456	128 272	44 128
<i>Metridia lucens</i> VI	48	1088	64
Station	C		
Hauling depth (m)	0-500	500-750	750-1000
Wire angle	32°		
Location	30°-00' N 144°-00' E		
Date	July 14th, 1965		
Time	11 ^h 30 ^m — 12 ^h 31 ^m		
<i>Calanus cristatus</i> VI V•IV	— —	— —	— 1
<i>C. plumchrus</i> VI V•IV	— —	— —	2 —
<i>Eucalanus bungii</i> VI <i>bungii</i> V•IV	— —	— —	— —
<i>Metridia lucens</i> VI	—	—	1

The roman numerals denote the stage of copepodids.

Table 5. Number of individuals (per a haul) of four boreal copepods obtained by sectional haul at the St. D.

Station	D			
Hauling depth (m)	0-120	120-250	250-500	500-750
Wire angle	14°		15°	
Location	38°-06' N 194°-49' E			
Date	Sep. 28th, 1963			
Time	14 ^h 30m		14 ^h 55m	
<i>Calanus cristatus</i> VI V•IV	— —	— —	— —	— 4
<i>C. plumchrus</i> VI V•IV	— —	— —	— 2	7 2
<i>Eucalanus bungii</i> VI <i>bungii</i> V•IV	— —	— —	2 40	17 65
<i>Metridia lucens</i> VI	—	3	42	4

The roman numerals denote the stage of copepodids.

(1) The Japanese Expeditions of Deep Sea

The four cold water species were collected abundantly at the Sts. A 3 and G 207 compared with other stations, though there was relatively great difference in the number of individuals between the two stations. The numbers of *Calanus plumchrus* and *Metridia lucens* diminished at the St. E 2, and the remained two species were reduced at the St. F 23. At the St. F 23, *Metridia lucens* was distributed in the layer from 500 to 1000 m depth, and *Calanus cristatus* and *Eucalanus bungii bungii* were located in the layers from 500 to 3000 m depth. Wire angle of a net being taken into consideration, the lower limit of their distribution may become shallower fairly, namely, it is presumed to be at about 2000 m depth. Besides, *Eucalanus bungii bungii* was occurred in the layer from 500 to 1000 m depth at the Sts. I 2 and I 23 and in the layers from 500 to 1800 m depth at the St. I 4. *Calanus plumchrus* was occurred in the layer from 500 to 1000 m depth at the St. I 4. Their number of individuals, however, was scarce, being less than 3.

(2) Other deep-sea plankton collections

The four species were collected abundantly at the Sts. A and B. Population

was concentrated in the layer from the surface to 500 m depth at the St. A and in the layer from 500 to 1000 m depth at the St. B. Adult of *Calanus cristatus* was entirely absent above 500 m, and the number of individuals increased in the layer from 750 to 1000 m depth. Referring to the previous data on the vertical distribution of this species, the upper limit seems to be below 150 m depth (ANRAKU, 1954). As to *Calanus plumchrus* and *Eucalanus bungii bungii*, adults were distributed from the surface to 1000 m depth, however, the number of individuals was extremely scarce in the layer above 500 m depth. It is reported that the adults of the two species have been collected above 150 m depth (FURUHASHI, 1952; MARUMO *et al.*, 1958; MINODA, 1958). Adult of *Metridia lucens* was abundant in the layer from 500 to 750 m depth, though it is a few in the other layers. At the St. D, the three copepods excluding *Metridia lucens* were distributed below 250 m depth having the smaller number of individuals. *Metridia lucens* was distributed below 120 m depth and had the center of population in the layer from 250 to 500 m depth. Its number was reduced in the same fashion as other species. At the St. C, three copepods excluding *Eucalanus bungii bungii* were distributed below 750 m depth, though the number of individuals decreased, being 4 in all.

4. Discussion

БЕКЛЕМИШЕВ (1961) has shown that *Calanus cristatus*, *C. plumchrus* and *Eucalanus bungii bungii* had their southern limit at about 35°N in the western North Pacific on the basis of the plankton samples taken in the layer from 100 to 500 m depth. In the present survey, it is clarified that the typical boreal copepods such as *Calanus cristatus*, *C. plumchrus*, *Eucalanus bungii bungii* and *Metridia lucens* are spread furthermore to the southern part along the Japan Trench. In that area, the center of the vertical distribution of these species is recognized at about 1000 m depth by the sectional collections, although their numbers decreased conspicuously in comparison with them in the north of the Oyashio front, and its depth coincides closely with it of the Subarctic Intermediate Water. At any rate, the above fact indicates that there is a southward flowing water having a origin in the Oyashio cold water in that depth.

Up to now, the collections of deep-sea plankton are very scarce in the sea area south of Japan. FURUHASHI and MATSUDAIRA (1960) have made the deep-sea plankton collection at two stations in the west of the Ogasawara Submarine Ridge, and FURUHASHI (1961) have made at two stations in the west and at a station in the east of the Ridge. Then, they reported that in the east of the Ridge, *Calanus*

cristatus and *Eucalanus bungii bungii* were taken, but in the west of it, any Oyashio cold water copepods were not observed. In this survey, *Calanus plumchrus* and *Metridia lucens* were added anew in the east of the Ridge. Moreover, at the St. I4 located on the Ridge, *Calanus plumchrus* and *Eucalanus bungii bungii* occurred, and at the St. I23 located in the west of the Ridge, *Eucalanus bungii bungii* occurred. This fact may indicate that there is the south-westward flow across the Ridge. Actually, NAN'ITI *et al.*, (1964) ascertained that there were a feeble south-westward and north-eastward currents in the 1000 m depth at the point east of the St. I4.

Recently, MARUMO (personal communication) suggested that *Calanus cristatus* was obtained with a relatively large number of individuals in the westside of the Ridge on the recent cruise of R. V. Tansei-Marui of the Ocean research Institute, Tokyo University. FURUHASHI (1965), also, reported that *Metridia lucens* have been found for the first time in the Enshu-Nada, off coast of the southern Japan. In conclusion, we might state that much more deep-sea plankton collections would be necessary to clarify the detailed movement of the water originating in the Oyashio cold water.

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A Numerical Computation of the Storm Surge of Hurricane Carla 1961 in the Gulf of Mexico*

By

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Abstract: The storm surge of Hurricane Carla 1961 in the Gulf of Mexico is computed by numerical integration of the dynamical equations. Input data consist of sea-level distributions of observed wind and pressure at six-hour intervals throughout the three-day period selected for the study.

The computation is carried out in two steps. The first is on a rough mesh (mesh length 48 nautical miles) for the entire Gulf; bottom friction is disregarded here. The second is a fine-mesh computation (mesh length 9.6 nautical miles) on the northwestern continental shelf, where the storm surge was most predominant. Quadratic bottom friction is assumed in the latter step.

The fine-mesh region is chosen so as to give good resolution in the 100-mile neighborhood of Galveston, Texas, which was directly struck by the Hurricane Carla in 1961. Computed variations of sea level are in good agreement with tide-gage observations in this region.

1. Introduction

Studies of numerical storm-surge prediction have been pursued actively during the past ten years. One of the first efforts along this line was made by KIVISILD (1954), for wind-tide computation on Lake Okeechobee. He made the computation by hand, but the use of the high-speed electronic computer soon accelerated the development of this field.

Later, HANSEN (1956) computed the storm surge on the North Sea induced by the 'Holland-Orken' of 1953, and PLATZMAN (1958) computed the surge of Lake Michigan accompanied by the moving squall line of 26 June 1954. From the

* This investigation was made while the writer was on leave from the Japan Meteorological Agency and serving in the University of Chicago (1962-63) and the U. S. Weather Bureau (1963-64). The work in Chicago was supported through funds provided by the U. S. Weather Bureau.

standpoint of future investigation, the important feature that these works have in common is the principle of numerical integration of the governing dynamical equations.

Subsequently, many people have been interested in this problem. Our knowledge has been extended and modified, for example, through the works of SVANSSON (1958), FISCHER (1959), UUSITALO (1960), HENNING (1962), PLATZMAN (1963), and others.

However, serious difficulties still remain for the application of this method to storm surges induced by tropical cyclones. This is chiefly because the precise pattern of the wind and pressure disturbances is very difficult to measure or evaluate in these cyclones, especially on an "operational" basis. MIYAZAKI *et al.* (1961-62) treated typhoon surges along the Japanese coast by assuming model patterns of the atmospheric-disturbance field, but it is not certain that these patterns are good representations over the sea because of the lack of observations.

Fortunately, a sequence of meteorological data and analyses were available during the passage of Hurricane Carla 1961 on the Gulf of Mexico, including 6-hourly maps of sea-level pressure and wind, which were analysed and prepared by the Hydrometeorological Section of the U. S. Weather Bureau. One of the aims of the present numerical experiments is to obtain better agreement between computed and observed hurricane storm surges by using these more accurate meteorological data.

Besides the rough-mesh computations for the entire Gulf, we make detailed computations with finer meshes on the northwestern continental shelf. The shelf is so wide in this neighborhood (about 100 miles in the average), and there are ample possibilities of the rapid storm-surge development on this area.

As for the finite-difference scheme, we use an elementary staggered grid system in which volume-transport components are given at odd time steps and surges are given at even time steps, respectively (cf. Figure 6). Equations are basically linear except for friction terms. The method used requires the implicit representations of Coriolis and friction terms in order to maintain the computational stability. However, the iteration technique is not necessary in our case, since prediction equations are easily derived from the primitive finite-difference equations.

Of course, there will still remain some difficulties which are derived from uncertainties of the surface- and bottom-stress laws, from truncation errors of the finite-difference scheme, including boundary conditions, and from various

other sources. The writer hopes that these difficulties will be removed one by one through continuous modification and repetition of numerical and experimental works.

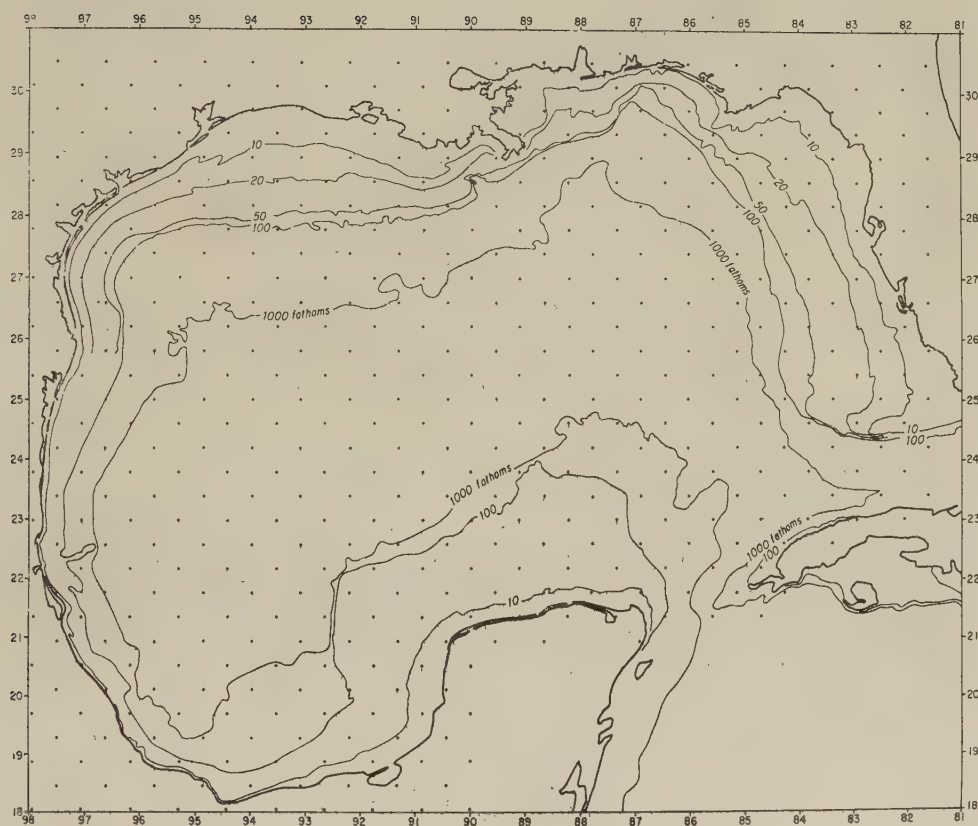


Figure 1. Bottom topography of the Gulf of Mexico.



Figure 2. The track of Hurricane Carla, 1961.

2. Basic equations and conditions

a) Dynamic equations of storm surge

We are based on the classic equations of EKMAN (1905), which were introduced for problems of the wind-induced motion on the rotating earth. Taking the z coordinate vertically downwards from the mean sea-surface, and x, y coordinates in a horizontal plane, the momentum and mass equations are

$$\left. \begin{aligned} \frac{\partial u}{\partial t} &= fv - g \frac{\partial h}{\partial x} + \frac{\partial}{\partial z} \left(\nu \frac{\partial u}{\partial z} \right), \\ \frac{\partial v}{\partial t} &= -fu - g \frac{\partial h}{\partial y} + \frac{\partial}{\partial z} \left(\nu \frac{\partial v}{\partial z} \right), \\ \frac{\partial h}{\partial t} &= -\frac{\partial M}{\partial x} - \frac{\partial N}{\partial y}, \end{aligned} \right\} \quad (2.1)$$

in which u, v is the horizontal velocity, g is gravity, h the elevation of free surface from its mean position, f the Coriolis parameter, ν the kinetic eddy viscosity, and $S(M, N)$ the horizontal volume transport per unit width from the free surface to the bottom.

From first two equations of (2.1), we easily obtain following equations of the volume transport

$$\left. \begin{aligned} \frac{\partial M}{\partial t} &= fN - gD \frac{\partial h}{\partial x} + \frac{1}{\rho} (\tau_s^x - \tau_b^x), \\ \frac{\partial N}{\partial t} &= -fM - gD \frac{\partial h}{\partial y} + \frac{1}{\rho} (\tau_s^y - \tau_b^y), \end{aligned} \right\} \quad (2.2)$$

where D is the water depth, ρ the sea-water density, and $\tau_s^x, \tau_s^y; \tau_b^x, \tau_b^y$ are components of the tangential stresses exerted on the sea-surface and bottom, respectively.

(2.2) holds when the normal stress on the sea surface is constant. If the atmospheric pressure exerts the variable normal stress on the free surface, (2.2) is slightly rewritten as

$$\left. \begin{aligned} \frac{\partial M}{\partial t} &= fN - gD \frac{\partial}{\partial x} (h - h_0) + \frac{1}{\rho} (\tau_s^x - \tau_b^x), \\ \frac{\partial N}{\partial t} &= -fM - gD \frac{\partial}{\partial y} (h - h_0) + \frac{1}{\rho} (\tau_s^y - \tau_b^y) \end{aligned} \right\} \quad (2.3)$$

by introducing the hydrostatic sea-surface elevation h_0 that corresponds to the departure of atmospheric pressure from a horizontal average.

(2.3) and the third equation of (2.1) are the basic equations which we are based on.

b) Initial and boundary conditions

Initial conditions are

$$M = N = h = 0 \quad \text{at} \quad t = 0. \quad (2.4)$$

The time origin is taken at 0600 Z, September 9, 1961. The hurricane center was about 400 nautical miles away from the Texas-Louisiana coast at that time. So influences of atmospheric disturbances were supposed to be small in this area.

The boundary condition at the coast is

$$S_n = 0. \quad (2.5)$$

where the suffix n denotes the transport component normal to the coast. It should be noted that the tangential component of volume transport also will be equal to zero when the depth vanishes at the coast.

Now, since the Gulf of Mexico is not a completely closed basin, the condition at the mouth should be taken into account in addition to the coastal condition (2.5). In the present case (cf. Figure 2), the hurricane center moved towards northwest from the Yucatan Strait to the middle of the Texas coast, and neither

winds nor atmospheric pressure changes were large in the neighborhood of the mouth except for the very early stage. Therefore, it seems likely that the condition

$$h=0 \quad \text{at the mouth} \quad (2.6)$$

will not cause significant errors for storm-surge phenomena in the northwestern part of the Gulf.

We are obliged to add another boundary condition

$$\frac{\partial S_n}{\partial n}=0 \quad (2.7)$$

along the southern edge of the fine-mesh computing area (cf. Figure 3). This edge line is taken sufficiently south from the region under consideration, but it crosses the continental shelf from the coast. The rough mesh computations for the entire Gulf (mesh size 48 naut. mi.) are not enough to estimate the boundary condition along this line because of the poor resolution on the shallow sea. (2.7) means that the incoming current across this line does not affect the continuity condition.

c) *Surface and bottom stresses*

The surface stress in equations (2.3) will be represented in the form

$$\tau_s = \rho_a \gamma_s^2 |\mathbf{w}| \mathbf{w}, \quad (2.8)$$

where $\mathbf{w}$ denotes the wind velocity at the anemometer height, ρ_a is the atmospheric density, and γ_s^2 is a non-dimensional constant. We take $\gamma_s^2 = 2.2 \times 10^{-3}$ and $k \equiv \rho^{-1} \rho_a \gamma_s^2 = 2.5 \times 10^{-6}$, respectively.

It is well known that there are many discussions of the formula (2.8). WILSON (1960), for example, collected nearly fifty γ_s^2 -measurements made by various methods; for strong winds the value ranged from 1.5×10^{-3} to 4.0×10^{-3} , with an average of 2.4×10^{-3} . However, effects of the bottom stress are included in some of above data which are given from the surface tilt observations. This is the reason why we assume a smaller γ_s^2 -value than his average.

This coefficient depends upon the state of the sea surface, as well as upon thermal stability of the lower atmosphere. However, at the present time, refinements of (2.8) are not worth-while since the measurement or evaluation of wind velocity in a hurricane system is still very difficult and subject to large errors.

As for the bottom stress, an approximation developed by REID (1957) should be applicable on a shallow sea. This is given by

$$\tau = \rho \gamma_b^2 |\mathbf{v}| \mathbf{v} - 5 \gamma_b \tau_s, \quad (2.9)$$

where γ_b is a constant, and $\mathbf{v}$ is the depth-mean current velocity.

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We shall use (2.9) on the continental shelf where the depth is than about 100 fathoms. For greater depths, where (2.9) no longer holds in general, we shall ignore the bottom stress and take

$$\tau_b = 0 \quad (2.10)$$

in place of (2.9).

3. Numerical procedure

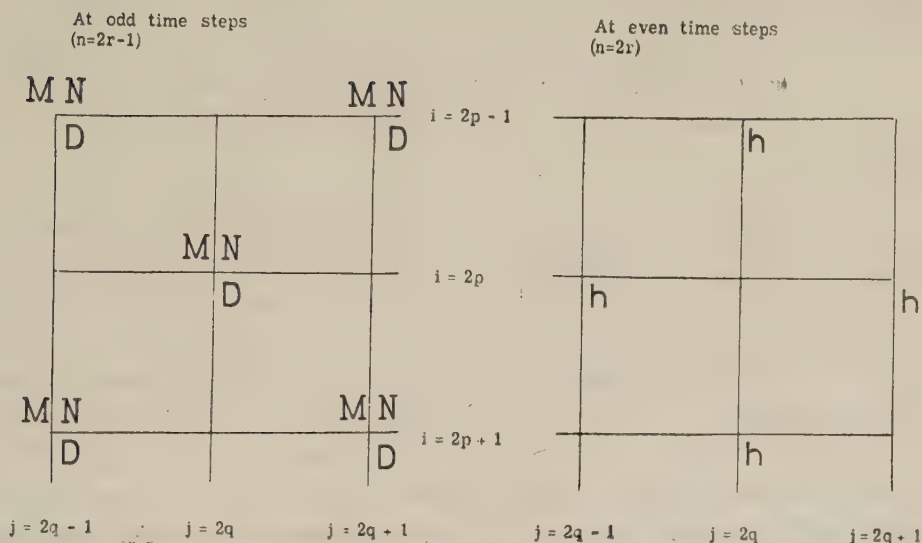
a) Finite-difference equations

By using (2.1), (2.3), (2.8) and (2.9), our basic equations are easily re-written as follows;

$$\left. \begin{aligned} \frac{\partial M}{\partial t} &= fN - gD \frac{\partial}{\partial x}(h - h_0) + T_x - GM, \\ \frac{\partial N}{\partial t} &= -fM - gD \frac{\partial}{\partial y}(h - h_0) + T_y - GN, \\ \frac{\partial h}{\partial t} &= -\frac{\partial M}{\partial x} - \frac{\partial N}{\partial y}, \end{aligned} \right\} \quad (3.1)$$

in which

$$\left. \begin{aligned} T_x &\equiv (1 + 5\gamma_b) \tau_s^x / \rho, \\ T_y &\equiv (1 + 5\gamma_b) \tau_s^y / \rho, \\ G &\equiv \gamma_b^2 \sqrt{M^2 + N^2} / D^2. \end{aligned} \right\} \quad (3.2)$$



For the purpose of numerical computation, (3.1) should be rewritten by finite-difference equations by using certain computational scheme.

In the following, we shall use a grid (i, j, n) which corresponds to the co-ordinates (x, y, t) in the differential equations. We shall adopt the simple central-difference scheme in which

- (i) volume-transport components are computed at odd n and even $(i+j)$ values, and
- (ii) elevations are computed at even n and odd $(i+j)$ values, respectively (cf. Figure 6).

In this scheme we have no transport values for even n . These values, which are necessary in the computation of Coriolis terms and of the friction terms, are defined by the average of transport at the odd- n steps immediately before and after each even n . In the friction terms, which are proportional to $|S|S$, we need to use this method only for the evaluation of S ; the factor $|S|$ is evaluated at the odd n , one step before each even n . The above processes are necessary in order to avoid computational instability.

If we introduce new variables

$$M' \equiv M \frac{\Delta t}{\Delta s}, \quad N' \equiv N \frac{\Delta t}{\Delta s} \quad (3.3)$$

in place of M and N , finite-difference equations will be easily written in the form (for odd n and even $i+j$)

$$\left. \begin{aligned} (M')_n^{i,j} &= A(M')_{n-2}^{i,j} + 2B(N')_{n-2}^{i,j} + CX_{n-1}^{i,j} + BY_{n-1}^{i,j}, \\ (N')_n^{i,j} &= -2B(M')_{n-2}^{i,j} + A(N')_{n-2}^{i,j} - BX_{n-1}^{i,j} + CY_{n-1}^{i,j}, \\ h_{n+1}^{i+1,j} &= h_{n-1}^{i+1,j} - (M')_n^{i+1,j} + (M')_n^{i,j} - (N')_n^{i+1,j+1} + (N')_n^{i+1,j-1}, \end{aligned} \right\} \quad (3.4)$$

in which

$$\left. \begin{aligned} A &\equiv \frac{1 - (\beta_{n-2}^{i,j})^2 - \alpha_j^2}{(1 + \beta_{n-2}^{i,j})^2 + \alpha_j^2}, \quad B \equiv \frac{\alpha_j}{(1 + \beta_{n-2}^{i,j})^2 + \alpha_j^2}, \quad C \equiv \frac{1 + \beta_{n-2}^{i,j}}{(1 + \beta_{n-2}^{i,j})^2 + \alpha_j^2}, \\ \alpha_j &\equiv \frac{1}{2} f_j \Delta t, \quad \beta_{n-2}^{i,j} \equiv \frac{1}{2} G_{n-2}^{i,j} \Delta t, \\ X_{n-1}^{i,j} &\equiv (T_x)_{n-1}^{i,j} - gD^{i,j} \left(\frac{\Delta t}{\Delta s} \right)^2 \cdot [h_{n-1}^{i+1,j} - h_{n-1}^{i-1,j} - (h_0)_{n-1}^{i+1,j} + (h_0)_{n-1}^{i-1,j}], \\ Y_{n-1}^{i,j} &\equiv (T_y)_{n-1}^{i,j} - gD^{i,j} \left(\frac{\Delta t}{\Delta s} \right)^2 \cdot [h_{n-1}^{i,j+1} - h_{n-1}^{i,j-1} - (h_0)_{n-1}^{i,j+1} + (h_0)_{n-1}^{i,j-1}]. \end{aligned} \right\} \quad (3.5)$$

Strictly speaking, the Gulf of Mexico is too large to permit us to ignore variations of map scale. Since we have adopted a Mercator projection, and have drawn a grid of equal-interval ΔS on this map, the true mesh size $\Delta S'$ is given by

$$\Delta S' = \Delta S \cdot \frac{\cos \phi}{\cos \phi_0}, \quad (3.6)$$

where ϕ is the latitude and ϕ_0 (in our case 25°N) is a standard latitude.

Accordingly, formulas (3.3)~(3.5) should be modified by using $\Delta S'$ instead of ΔS . However, multiplicative operations can be avoided by introducing a new variable

$$h' \equiv h \frac{\cos \phi}{\cos \phi_0} \quad (3.7)$$

in place of h , and

$$D' \equiv D \left(\frac{\cos \phi}{\cos \phi_0} \right)^2 \quad (3.8)$$

in place of D . By these two substitutions, equations (3.4) can be used without changing any of other coefficients.

b) Errors from the coastal condition

No computational difficulties will arise from the coastal condition, since we adopted a zigzag boundary (Figure 3) with no depths. This means that we only use the continuity equation along the coast.

However, the zigzag boundaries introduce a series of artificial basins whose lateral size is ΔS . We must therefore insure that the maximum period $T_{\max}$ of artificial oscillations in these basins is shorter than the minimum period of physically significant events that accompany the storm surge, because coastal phenomena with period comparable to or shorter than $T_{\max}$ presumably are distorted in the computation with zigzag boundary.

In our case, $T_{\max}$ is approximated by

$$T_{\max} = \sqrt{\frac{2}{g D_{\min}}} \pi \Delta S, \quad (3.9)$$

where $D_{\min}$ is the minimum depth at mesh points adjacent to the coast. The minimum time scale which we must take into account will be about 20 hours for the large-scale phenomena, and about 5 hours for phenomena on the north-western continental shelf. The former value is fixed by the time gap between two oppositely-directed wind maxima of the moving hurricane, and the latter value corresponds to the period of the primary eastward-moving edge waves (c. f. REID, 1958) along the Texas-Louisiana coast. Any significant changes in the forcing terms will be of longer period than the former, and any significant oscillation on the northwestern continental shelf, such as the proper oscillations, westward edge waves, or inertial waves, have longer periods than the latter.

Meanwhile, our mesh-sizes are 48 (approximately) and 9.6 nautical miles for the rough-mesh and fine-mesh computations, respectively. The minimum mesh-point depth is 5 fathoms in both cases. Then, maximum noise periods will be 11.2 and 2.24 hours, respectively. They are fairly shorter than the minimum periods which we want to take into account in these computations.

c) *Stability of the computational scheme*

For the typical solution of the type

$$\left. \begin{aligned} M' &= M_p \exp(mt + i\kappa x + tly), \\ N' &= N_p \exp(mt + i\kappa x + ily), \\ h' &= h_p \exp(mt + i\kappa x + ily), \end{aligned} \right\} \quad (3.10)$$

(3.4)¹ becomes

$$\left. \begin{aligned} (A - \lambda)M_p + 2BN_p + (C\mu + B\nu)i\phi h_p &= 0, \\ -2BM_p + (A - \lambda)N_p + (B\mu - C\nu)i\phi h_p &= 0, \\ \mu i\lambda M_p - \nu i\lambda N_p N_p + (1 - \lambda)h_p &= 0, \end{aligned} \right\} \quad (3.11)$$

in the absence of forcing terms, where

$$\lambda \equiv e^{2m\Delta t}, \quad \mu \equiv 2 \sin(\kappa \Delta s), \quad \nu \equiv 2 \sin(l \Delta s), \quad \phi \equiv gD' \left(\frac{\Delta t}{\Delta s} \right)^2 (\mu^2 + \nu^2), \quad (3.12)$$

and parameters A , B , and C are defined by (3.5).

(3.11) gives a cubic equation of λ , from which we obtain

$$\left. \begin{aligned} \lambda_1 &= \frac{1 - \beta}{1 + \beta}, \\ \lambda_{2,3} &= \frac{1 - \alpha^2 - \frac{\phi}{2} \pm \sqrt{(\beta^2 - 4\alpha^2) - (1 - \alpha^2)\phi + \frac{\phi^2}{4}}}{(1 + \beta)(1 + \alpha^2)}. \end{aligned} \right\} \quad (3.13)$$

For stability it is necessary and sufficient that $|\lambda_i| \leq 1$ ($i=1, 2, 3$)². Since $\beta \leq 0$, $|\lambda_1| \leq 1$ is always satisfied. $|\lambda_2| \leq 1$ and $|\lambda_3| \leq 1$ give $\phi \leq 4$. From (3.12) this leads to the condition

$$gD'_{\max} \left(\frac{\Delta t}{\Delta s} \right)^2 \leq \frac{1}{2},$$

or

$$\frac{\Delta s}{\Delta t} \geq \sqrt{2gD'_{\max}}, \quad (3.14)$$

since $\mu^2 \leq 4$ and $\nu^2 \leq 4$.

1. If the map scale is variable, h and D in (3.4) should be replaced by h' and D' , respectively (see (3.7) and (3.8)).
2. See PLATZMAN (1963), p. 37.

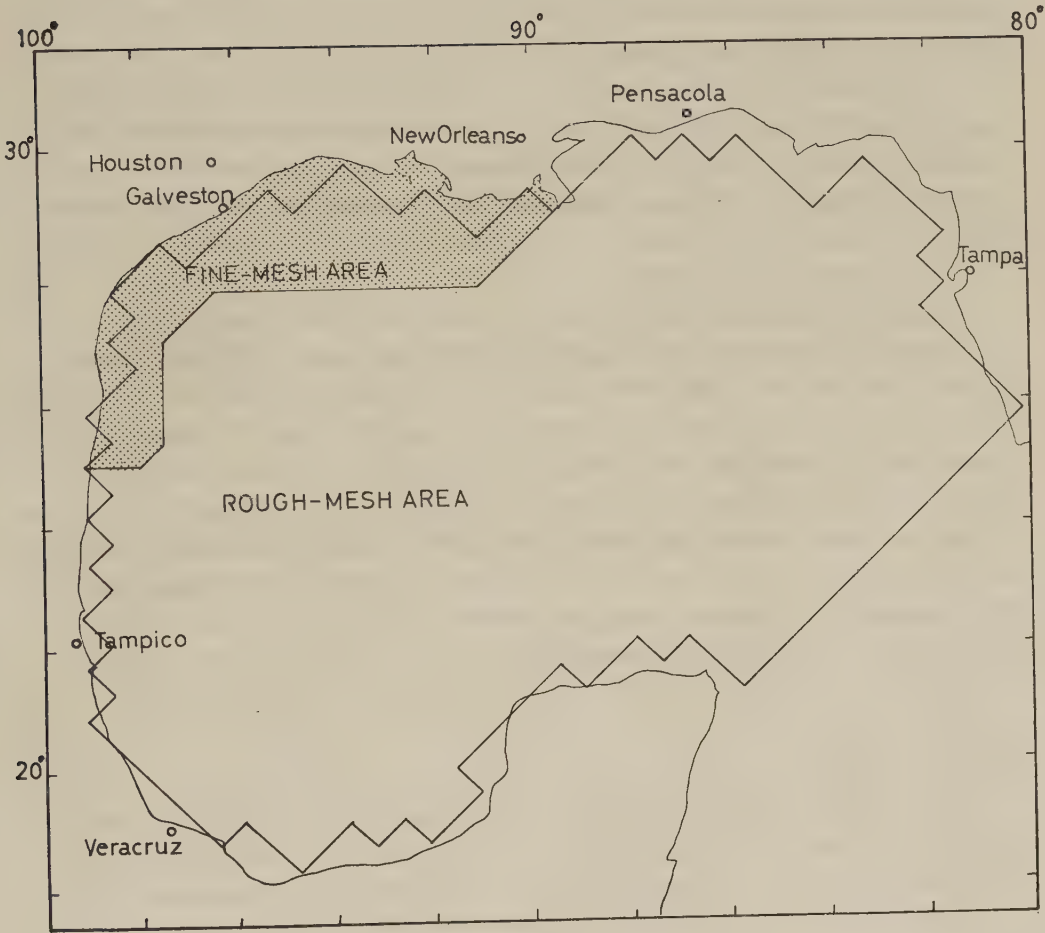


Figure 4. Computational area.

(3.14) represents the necessary and sufficient condition for the computational stability. $D'_{\max}$ is the maximum value of the modified mesh-point depth.

d) *Plan of the numerical computation*

As previously noted, we make two phases of the computation. One is the rough-mesh computation for the entire Gulf, and another is the fine-mesh computation for its northwestern part.

(i) Rough-mesh computation

The purposes of this computation are to obtain the rough features of the phenomena over the whole Gulf, and to obtain boundary conditions for computation in the coastal area. Bottom friction is disregarded here. The mesh is equal-interval on a Mercator map, and the true mesh size is 48 n.mi. at the standard latitude 25°N .

(ii) Fine-mesh computation

The northwestern continental shelf from Texas to Louisiana ($D \lesssim 85$ fm.) is considered in this phase of the computation. The mesh size is 9.6. n.mi. Reid's approximate law for bottom stress is used. At the seaward edge, the elevation h given from the rough-mesh computation is adopted, since we expect only small influences from the outer deeper area (Appendix II). Along the southern edge, we adopted the condition (2.7).

The computing regions for these two phases are shown in Figure 4. The time interval Δt is 5 minutes in both of these computations. The time origin of the computation is taken at 0600 Z, September 9, 1961. More details on the computing plan are explained in Appendix I.

4. *Meteorological data*

a) *Hurricane Carla, 1961*

The track and the central pressures of Carla are shown in Figure 2. This hurricane was remarkable for its slow speed of movement. The center was in Yucatan Strait at 1500 Z, September 7, with the surface pressure 970 mb. As it deepened, it moved very slowly to northwest. The central pressure reached its minimum 936 mb at 1200 Z, September 10. The location then was still 200 n. mi. SSE of Galveston, Texas.

The center of hurricane reached the coast approximately at 2000 Z September 11, near the east end of Matagorda Island, Texas; the central pressure was about 938.5 mb at that time.

The main speed during passage across the Gulf of Mexico was not more

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than about 7 knots. The maximum wind speed during this passage was evaluated to be 85-95 knots. The scale of this hurricane was fairly large, the maximum wind zone being about 50 n. mi. distant from the center.

b) Source of meteorological data

Basic meteorological data are available at the National Weather Record Center. However, we have used more convenient material which were already compiled and analyzed by the Hydrometeorological Section, U. S. Weather Bureau.

This Section made 6-hourly surface-pressure charts over the Gulf by using the observations and empirical relations between the pressure and the central distance. Isotach charts also are available, which are based upon observations and numerical computations that involve the pressure patterns³.

c) Pressure data

Surface-pressure charts are available for the three-day period 1200 Z September 12. Each chart shows isobars of surface pressure at 8-mb intervals. It was not difficult to read the required mesh-point values from these original charts.

d) Wind data

Isotach charts also are available for the same period, but those in the latter part of this period were not adequate for the analysis explained in (e) below, since the hurricane center was far from the computing area. The last available chart was that of 0600 Z September 12.

Isotachs are drawn at 10-knot intervals on these charts. The charts also include observed wind records, and computed wind directions. By using these data, we drew isogons of the inflow angle to the hurricane center at 10-degree intervals. One of these charts is shown in Figure 5.

Now, it is apparent that these charts indicate large local disturbances of small scale. The contribution from these local disturbances will not be of primary importance for the storm surge, but may be difficult to resolve by our finite-difference computations. Therefore, it is desirable to remove high wave-number contributions from the wind stress. For this purpose. We make a polynomial smoothing of the wind stress.

e) Wind stress analysis

Let us consider the expansion

$$f(r, \theta) = \sum_{m=0}^{m_0} \sum_{s=0}^{m_0-m} P_m(r/a) (a_{ms} \cos s\theta + b_{ms} \sin s\theta) \quad (4.1)$$

3. See GRAHAM and HUDSON (1960), and MYERS and MALKIN (1961).

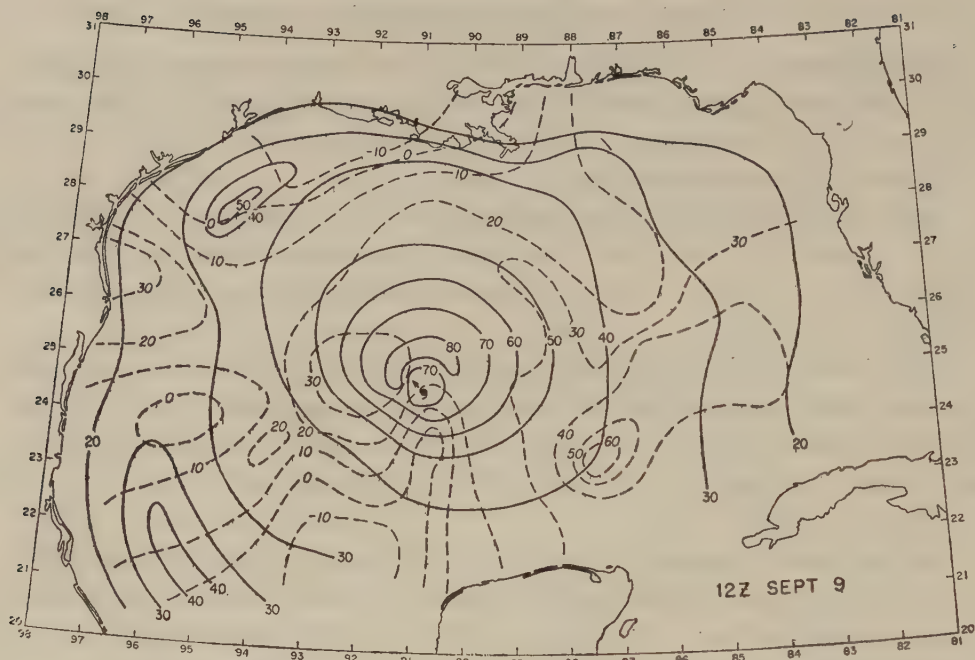


Figure 5. Example of an analyzed wind map (12Z September 9, 1961).
 full lines—isotaches (in kt)
 dashed lines—contours of the equal inflow angle against the
 hurricane center

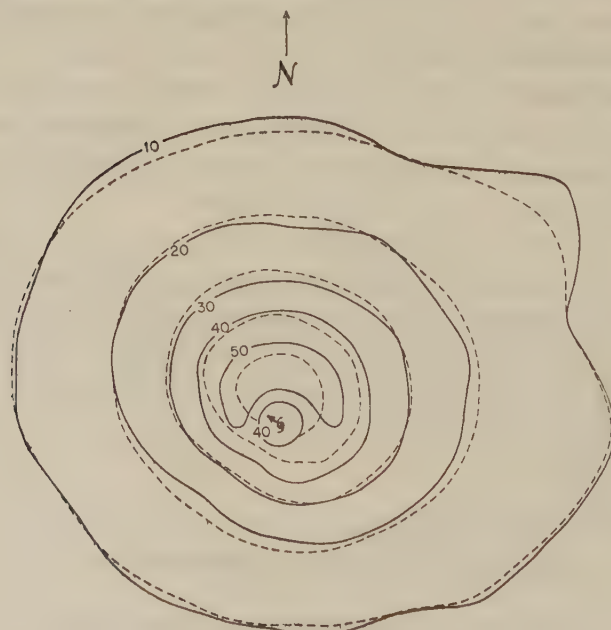


Figure 6. The observed (full-line) and reconstructed (dashed-line) contours of
 wind stress in dynes cm^{-2} at 12Z September 9, 1961. (Tangential com-
 ponent against the hurricane center).

in the region A where $0 \leq r \leq a$ and $0 \leq \theta \leq 2\pi$. In (4.1), $P_m(x)$ denotes the orthogonal polynomials that satisfy the condition

$$\left. \begin{aligned} \int_0^1 x P_m(x) P_n(x) dx &= 1 && \text{when } m=n \\ &= 0 && \text{when } m \neq n. \end{aligned} \right\} \quad (4.2)$$

Actual forms of $P_m(x)$ are

$$\left. \begin{aligned} P_0(x) &= \sqrt{2}; \quad P_1(x) = 4 - 6x; \\ P_2(x) &= 3\sqrt{6} \left(1 - 4x + \frac{10}{3}x^2 \right); \\ P_3(x) &= 8\sqrt{2} \left(1 - \frac{15}{2}x + 15x^2 - \frac{35}{4}x^3 \right); \dots \end{aligned} \right\} \quad (4.3)$$

and the relations

$$\left. \begin{aligned} \frac{1}{\pi a^2} \int_{A(r,\theta)} f(r,\theta) P_m(r/a) \cos s\theta dA &= 2a_{ms} \quad \text{for } s=0 \\ &= a_{ms} \quad \text{for } s \neq 0 \\ \frac{1}{\pi a^2} \int_{A(r,\theta)} f(r,\theta) P_m(r/a) \sin s\theta dA &= b_{ms} \end{aligned} \right\} \quad (4.4)$$

determine the coefficients a_{ms} and b_{ms} of equation (4.1).

The expansion (4.1) was used for the purpose of removing high wave-number portions from the original function $f(r,\theta)$. We applied this procedure for both tangential and radial wind-stress components, which are computed from the wind speed and the inflow angle to the hurricane center.

With $a=300$ n. mi., the coefficients a_{ms} , b_{ms} are computed for each chart, by means of numerical integration of (4.4) by Simpson's one-third rule. The mesh intervals for this purpose are 30 n. mi. (two end values and the central value are used for each interval) in the radial direction, and 40 degrees in the tangential direction, respectively.

Table 1. Some numerical values of A_{ms} and b_{ms} , the harmonics of wind stress, in dynes cm^{-2} .

(i) Tangential component, OOO September 10

m	a_{m0}	a_{m1}	a_{m2}	a_{m3}	b_{m1}	b_{m2}	b_{m3}
0	7.096	-0.346	-0.636	-0.126	-2.271	-0.414	-0.331
1	0.8395	-0.0900	-0.0331	—	-0.3167	-0.0137	—
2	0.0842	-0.0032	—	—	0.0063	—	—
3	0.0040	—	—	—	—	—	—

(ii) Tangential component, Ooz September 11

m	a_{m0}	a_{m1}	a_{m2}	a_{m3}	b_{m1}	b_{m2}	b_{m3}
0	7.647	-1.599	-0.545	-0.313	-1.919	-0.300	-0.223
1	1.2840	0.0734	-0.0513	—	-0.4240	-0.0049	—
2	0.0764	0.0040	—	—	0.0242	—	—
3	-0.0067	—	—	—	—	—	—

(iii) Radial component, Ooz September 10

m	a_{m0}	a_{m1}	a_{m2}	a_{m3}	b_{m1}	b_{m2}	b_{m3}
0	0.594	-1.266	0.993	0.172	-1.325	-0.496	0.270
1	0.4556	0.1213	0.0118	—	-0.3773	-0.0377	—
2	0.0319	0.0326	—	—	0.0127	—	—
3	-0.0009	—	—	—	—	—	—

After computing all coefficients for $m_0=3$, we recompiled them in (4.1) to obtain smoothed patterns of forcing functions. Table 1 shows some examples of the analyzed values, and Figure 6 shows an example of the comparison between the original and smoothed patterns. They are near coincidence except for a narrow region near the center.

5. Storm surge data

Due to Hurricane Carla, slow but severe sea-level rises were observed all along the Texas-Louisiana coast for almost six days from September 7 to 12. The tide gage of Galveston (Pleasure Pier) shows the maximum surge height of 8.8 ft at midnight of September 10. The maximum surge evaluated from the maximum-height observations, which include some wave effects, are about twice this value along the coast. The corresponding maximum value without the wave set-up would be about 10 ft or slightly more.

As for the tidal records, more than 20 records are available along the northern Gulf coast, where the storm surges were most dominant. Among these, we have reliable observations with precise tidal constants at the following stations:

1. Port Isabel, Texas
2. Port Aransas, Texas
3. Freeport, Texas

- | | | |
|---------------------------------------|---|----------------|
| 4. Pier 21, Texas | } | Galveston Area |
| 5. Pleasure Pier, Texas | | |
| 6. Fort Point, Texas | | |
| 7. Sabine Pass, Texas | | |
| 8. Bayou Rigaud, Louisiana | | |
| 9. Humble Oil Platform "A", Louisiana | | |
| 10. Pensacola, Florida | | |

At each of these stations, we have hourly values of observed sea level, and of the predicted astronomical tide (prepared by the Storm-Surge Project of the U. S. Weather Bureau). The residual values (observed minus predicted) approximately represent the storm surge. These data were available for the period September 6-13.

Besides these precise tidal data, we can also use other tide-gage records and observations of the highest sea level at many coastal points. These data have been summarized by HARRIS (1963).

6. Results of of computation

a) Storm surge at Selected stations

We have computed the storm surge at 18 stations along the Texas-Louisiana coast. Of these seven can be compared with the observed surge on the basis of tide-gage records. These stations are as follows:

Texas: Port Isabel, Poat Aransas, Pass Cavallo, Freeport, Galveston, Sabine Pass;

Louisiana: Humble Oil Platform (Bayou Rigaud).

Figure 7-13 shows the computed (thin lines) and observed (thick lines) variations at these stations.⁴

Now, some of the above stations face coastal inlets or lagoons, which are not considered in our present computations. Our computed results should correspond to values along the main open coast⁵. In this respect, we must first check the location of each station before making comparison of results.

-
4. At Pass Cavallo, we have no tidal records. The observed tide at Port O'Connor (above MSL) is shown by the dotted line (Figure 9).
 5. In later stages when inland flood water is abating, the sea level along the main coast can be affected by it near the mouth of inlets. However, the effect will be small since the volume of the outflow water is negligible against that of the continental shelf and if changes in forcing terms are small.

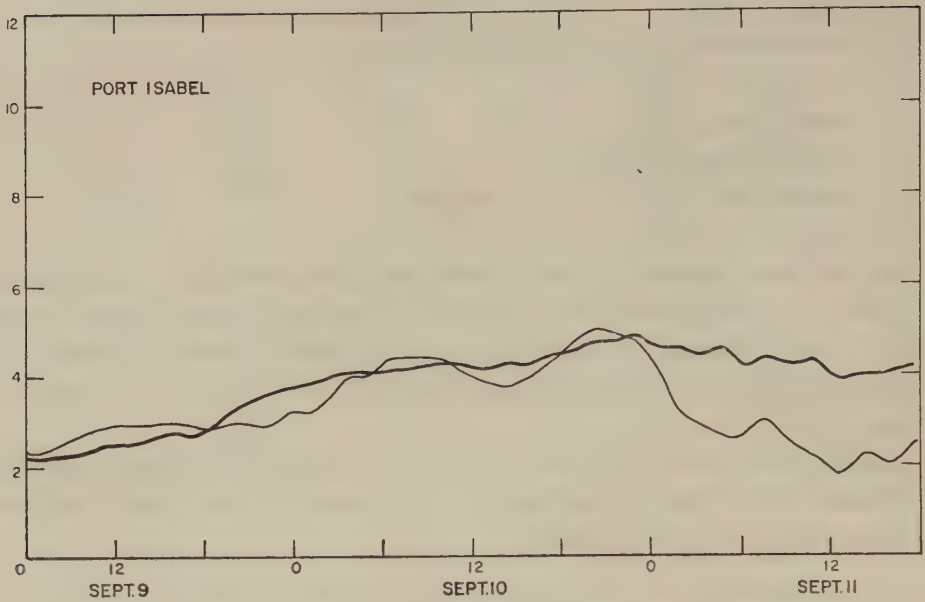


Figure 7. Observed (thick-line) and computed (thin-line) time variations of the storm surge at Port Isabel, Texas.

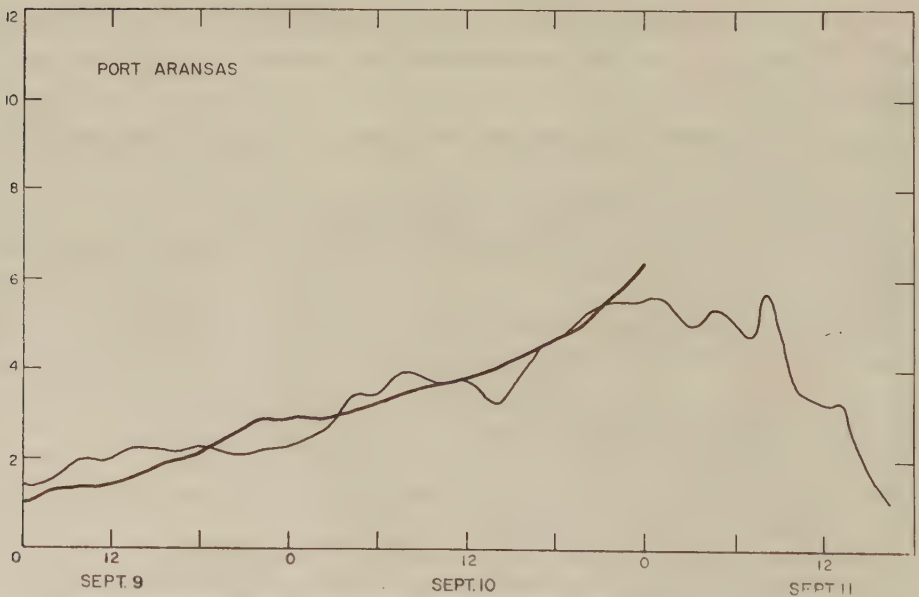


Figure 8. Observed (thick-line) and computed (thin-line) time variation of the storm surge at Port Aransas, Texas.

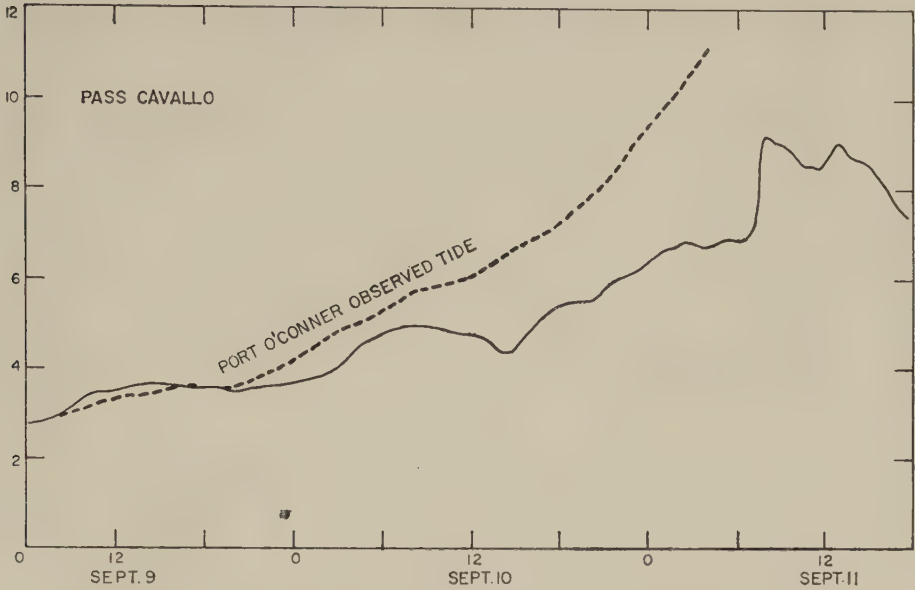


Figure 9. Computed surge at Pass Cavallo, Texas, and the observed tide at Port O'Connor, Texas.

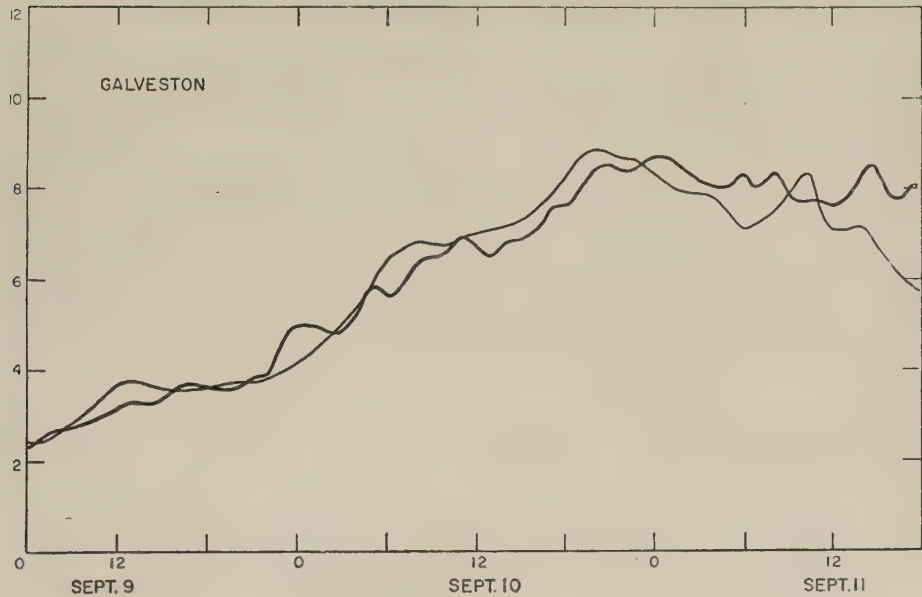


Figure 10. Observed (thick-line) and computed (thin-line) time variations of the storm surge at Freeport, Texas.

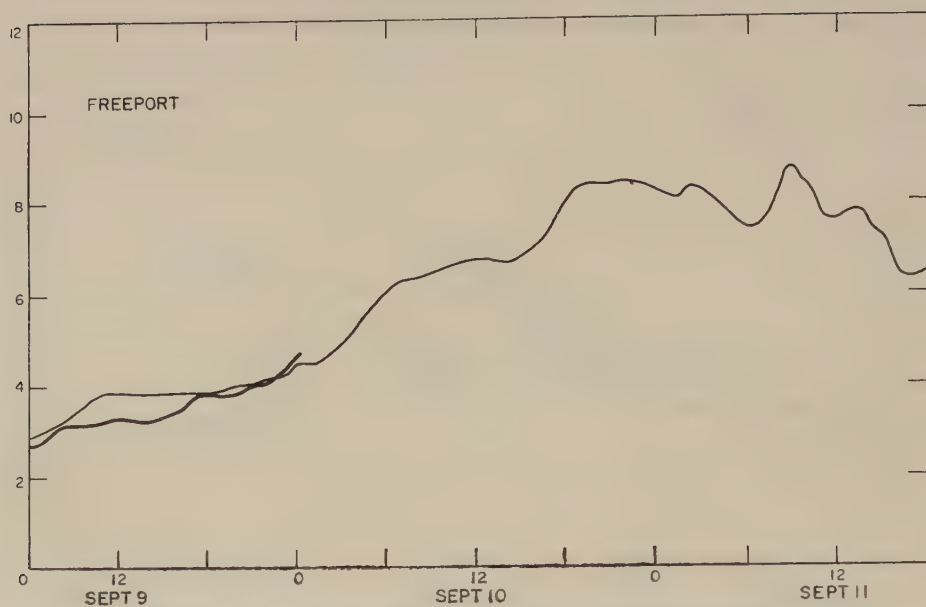


Figure 11. Observed (thick-line) and computed (thin-line) time variations of the storm surge at Galveston, Texas (Pleasure Pier).

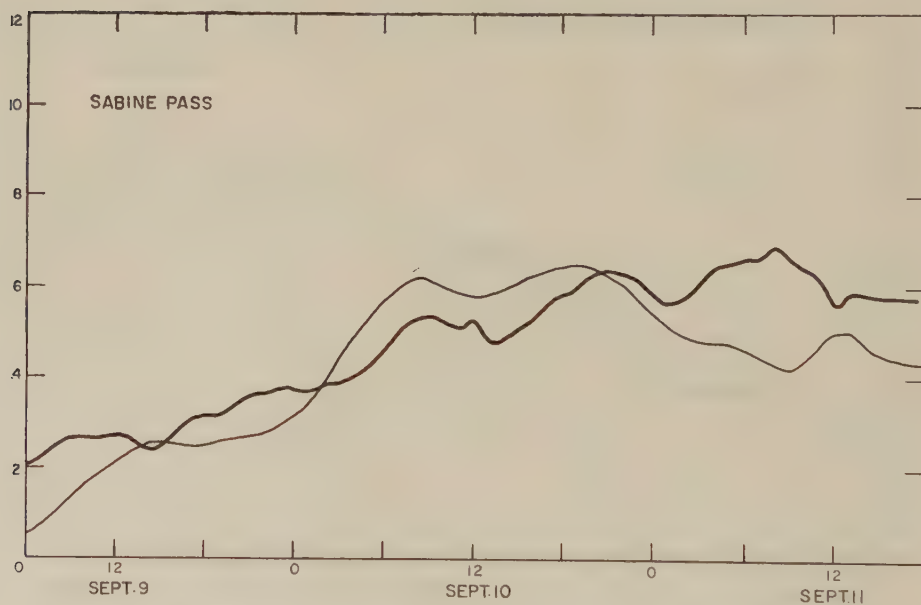


Figure 12. Observed (thick-line) and computed (thin-line) time variations of the storm surge at Sabine Pass, Texas.

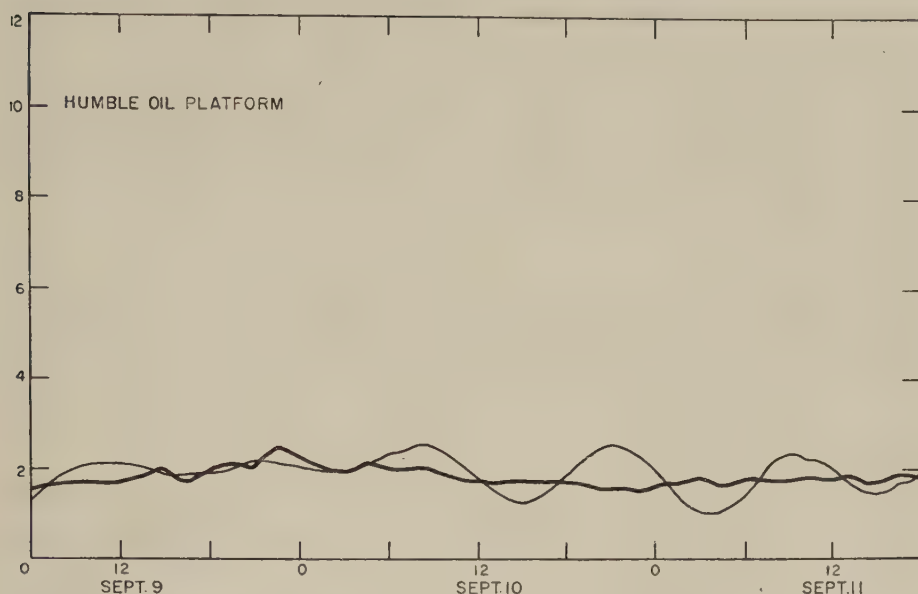


Figure 13. Observed (thick-line) and computed (thin-line) time variations of the storm surge at Humble Oil Platform, Louisiana.

Among them, Freeport, Galveston (Pleasure Pier), and Humble Oil Platform are facing the open sea. Port Aransas is not along the main coast, but is connected with the open sea through a short, wide, and deep channel (0.8 mi. $\times$ 0.2 mi. $\times$ 40 ft). Therefore, storm surges would not be so much different from those stations just outside the channel.

However, matters are somewhat different at other stations. Port Isabel and Port O'Connor are facing some nearshore lagoons, and Sabine Pass is located along a narrow pass connecting the open sea and a lagoon (Sabine Lake). Tidal retardations due to the inland water passage are listed in Table 2 for these three

Table 2. Tidal retardations of three inland stations against their open mouths.

Station	Time Lag (min.)		Ratio of the diurnal diurnal
	HW	LW	
Port Isabel	86	63	0.9
Port O'Connor	—	—	0.4
Sabine Pass	26	16	0.8

stations. All of the tabulated values are based on tidal constants at these stations and those at their open mouths (cf. Tide Tables 1965).

The tidal response is poor at Port O'Connor, and we can hardly fix the time lag of Lides. This would suggest less dependence of the storm tides on main-coast values. In contrast, observed storm surges at Port Isabel and Sabine Pass would surely be the measure of main-coast surges because of their high tidal responses.

Now, under these situations, computed surges in Figures 7-13 seem to be quite reasonable. Computed surges are in good agreement with observed surges at three west-side stations, Port Aransas, Freeport, and Galveston, so far as the available records are concerned. Observed records of Port Aransas and Freeport are not complete, only Galveston had a complete tide-gage record for the entire storm period.

Some differences are shown between computed and observed surges at Sabine Pass, but discrepancies are in a reasonable sense. Computed surge oscillations are somewhat larger and earlier than those observed in the earlier stages when we had inflow from the open sea, while the computed surge is lower than observed ones in the later stages when the flooded inland water was abating.

The situation is nearly similar at Port Isabel, but the above features of the computed surges are not so clear in the earlier stages perhaps due to the insufficient fetch in this neighborhood. Some correspondences are shown between the computed surges at Pass Cavallo and the observed tide at Port O'Connor, although there are many discrepancies.

At the Humble Oil Platform, where reconstructed winds were weak and fluctuating, 12-hour period oscillations with the maximum amplitude of 0.8 ft are present in the computed storm surge, but not in the observed one. They are probably due to the 6-hourly interpolation of the wind-stress values.

b) Horizontal patterns of the storm surge on the Gulf

Figure 14 (a-d) shows 12-hourly maps of storm-surge patterns over the whole area.

At 00Z and 12Z September 10 (Figure 14 a, b), a peak of 1-2 ft is present on the deep sea, in a position which corresponds roughly to the location of the hurricane center. This peak is due mainly to the direct effect of atmospheric pressure. At later stages, the peak seems to be merged in the high-surge area on the northwestern continental shelf. We also have a small high-surge area to the north of the Yucatan Peninsula.

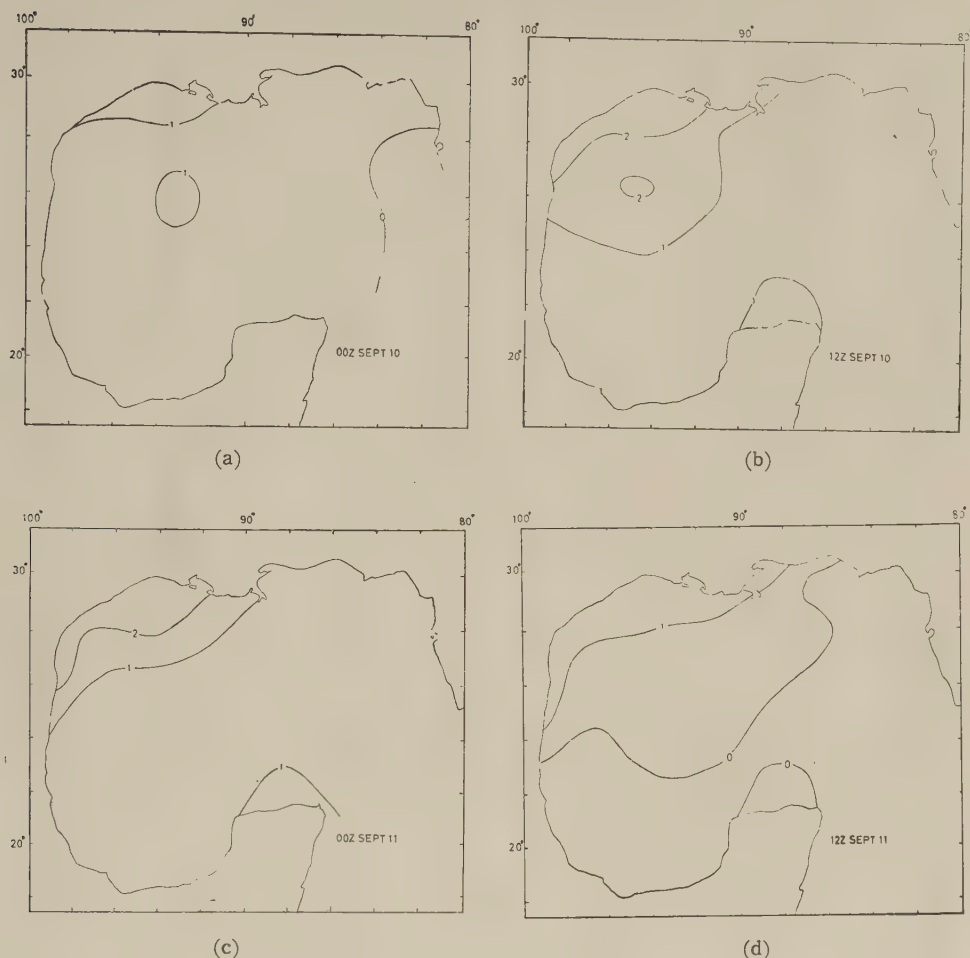


Figure 14. Horizontal patterns of the storm surge on the whole Gulf.

c) Horizontal patterns of storm surge on the northwestern continental shelf

Two typical patterns of the horizontal storm-surge distribution are shown in Figures 15 and 16. They are computed patterns at 00Z and 18Z September 11, 1961. In these figures, thick and thin lines indicate contours of the sea-surface elevation and the depth, respectively. Each arrow shows the distribution of reconstructed winds. The length of the arrow is taken to be proportional to the wind-speed square at that point.

(i) Horizontal pattern at 00Z September 11 (Figure 15)

The center of the hurricane was about 145 statute miles south of Galveston,

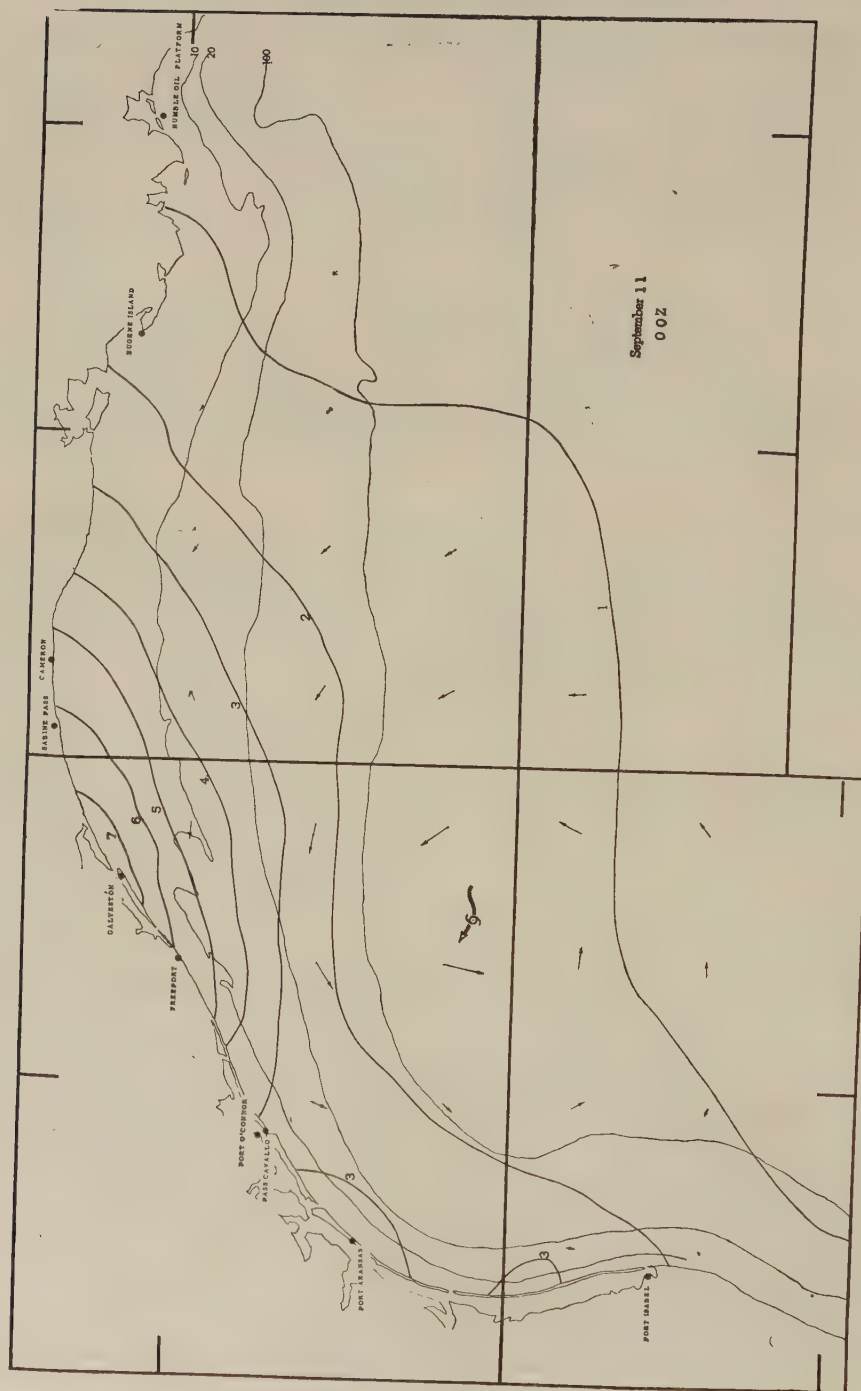


Figure 15. Horizontal pattern of storm surge on the northwestern continental shelf and its neighborhood at 00Z September 11, 1961 (Arrows show the wind velocity).

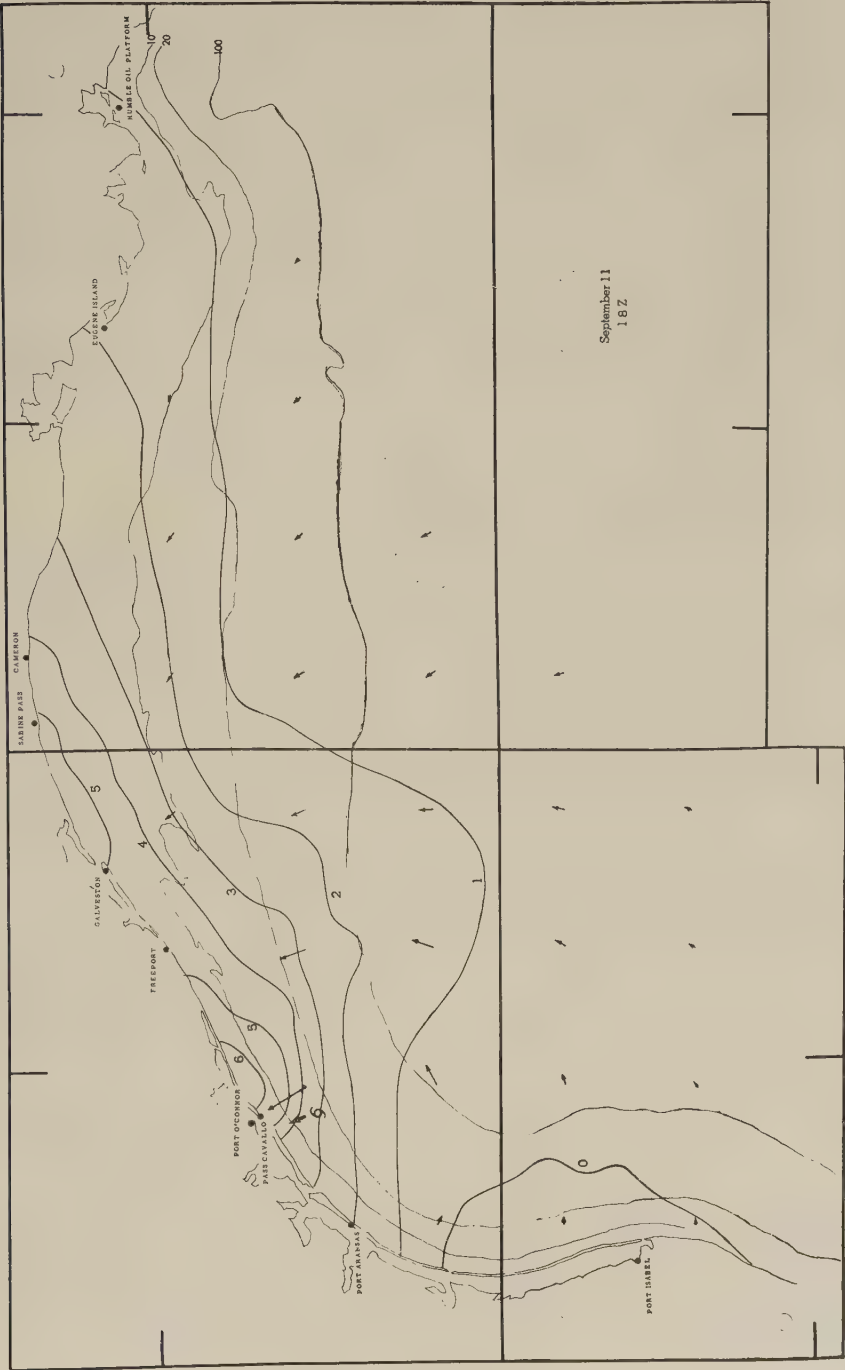


Figure 16. Horizontal pattern of storm surge on the northwestern continental shelf and its neighborhood at 18Z September 11, 1961 (Arrows show the wind velocity).

Texas. It was shifting towards the west with a slow speed of 9.5 knots. The meandering of its track as indicated in Figure 15 was remarkable. The highest wind near the center (10-minute average) was estimated to be about 100 knots.

Under these circumstances, strong easterly winds with some onshore components predominated for several hours on the wide continental shelf eastward from the Galveston area. To the west of this area, the shelf gradually becomes narrow and winds also had offshore components. Consequently, the water should be piled up strongly towards the Galveston area. Our computation agrees well with this theory, and the maximum value we get is about 8 ft.

Computed surges are higher than 1 ft in almost all part of the fine-mesh computing area. The 1-ft line is also extended to the deeper region in the central part of this area. We also have two secondary maxima along the western coast, but there are few observed data to verify them.

(ii) Horizontal pattern at 18Z September 11 (Figure 16)

The hurricane was only 30 miles offshore from Pass Cavallo, Texas, and moved towards NNW. Computed surges are more than 6 ft along the south shore of Matagorda Peninsula, which is located a little east of the hurricane track. Another maximum in the Galveston area still remains in this period, and the value is approximately 6 ft.

Negative surges are computed along the western coast in the front-left quadrant of the hurricane, but amplitudes are less than 1 ft.

It is to be noted that our computation shows the separated appearance of two major maxima. One appeared in the Galveston area when easterly onshore winds were mostly dominant on the wide continental shelf in the east of this area. The time was shortly after the period of Figure 15. Another maximum took place in the area of O'Connor accompanied by landfall of the hurricane. The time was shortly after the period of Figure 16. According to our results, the former maximum would be slightly larger than the latter.

d) *Currents associated with storm surge*

The wind-induced current is trivial except for the coastal region. On the northwestern continental shelf, the predominant direction of the current was south to west along the coast. The maximum speed was about 3 knots near the offward edge of the shelf, and about 2 knots near the coast. In later stages (18Z September 11, or latter), the current direction turned to north in the southern part of this area, and to east in some of deeper regions in the eastern part. The hurricane center was on land, or very close to the land at those times.

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Acknowledgements: This work was made when the writer served in the University of Chicago (1962-63) and the U. S. Weather Bureau (1963-64). The basic plan of the research was fixed in Chicago by the suggestion and guidance of Prof. G. W. PLATZMAN, and the final computation with some modifications was carried out in Washington, D. C., while he was associated with the Storm Surge Project of the U. S. Weather Bureau headed by Mr. D. Lee HARRIS. The writer wished to express his thanks to both of them.

He is also indebted to Dr. D. B. RAO, University of Chicago, and to Mr. Chester P. JELESNIANSKI, Storm Surge Project, for their kind advice and help. Useful advice from Dr. K. MIYAKODA, Dr. K. BRYAN, and Mr. N. A. PORE is also appreciated.

Appendix I

Description on the computer program

a) The running program

For convenience of computation, we ran three programs sequentially on the IBM 7030 (Stretch) computer.

The first program corresponds to the rough-mesh computation. Besides the usual output from this program, we keep 6-hourly mesh-point values of the disturbances, and also half-hourly values of elevation along the edge of the north-western continental shelf.

The second program starts immediately after the finish of the first program, and corresponds to the fine-mesh computation. The 12-hourly horizontal patterns are stored on the output tape. The elevation at selected stations are kept on the scratch tape and are re-arranged in the third program, which gives the remaining output data.

The speed of computation is 4 to 5 minutes of machine time per 24 hours of real time.

b) Input data

The input data consisted of the following.

(i) Computing region

For each i , the first and the last value of j are punched. They correspond to h -points next to the coast or the open edge.

(ii) Selected stations

The (i, j) coordinates of the selected point at which the storm-surge variations are to be given as output data.

(iii) Depth values

Depth values at each S -point (in fm).

(iv) Some constants

These define the period of the computations, coefficients of finite-difference equations, etc.

(v) Locations of the hurricane center

The (i, j) coordinates at each map time are given.

(vi) Pressure and wind-stress values

These data have been already explained in Section 4.

c) *Output data*

The output data consisted of the following.

(i) Twelve-hourly horizontal patterns of the flow and sea-surface elevation.

(ii) Half-hourly values of sea-surface elevation at 18 selected points along the Texas-Louisiana coast.

d) *Interpolation of wind and pressure*

Since the wind and pressure values are given only 6-hourly, we must make interpolations of these values in order to obtain mesh-point values at each time step. Interpolation is also necessary in order to obtain the h -values at the seaward edge in the fine-mesh computation. For these purposes we use the following methods.

(i) Pressure values

In the rough-mesh computation, pressure values are given at each h -point from 6-hourly charts. For the intermediate time steps, linear interpolation between 6-hourly values is used. In the fine-mesh computation, linear interpolation is also made with respect to the i, j coordinates.

(ii) wind-stress values

The coefficients a_{ms} , b_{ms} ($m, s \leq 3$) of tangential and radial wind stresses are computed at each 6-hour step. At these time steps, we computed the central distance r and the angle θ measured anti-clockwise from west at each S -point of the rough-mesh. Then the wind-stress components are computed by using the values of r , θ and a_{ms} , b_{ms} . At time steps intermediate between map times t and $t+6$ hr, we use quadratic interpolation that conserves the stress at t and $t+6$ hr, and the stress tendency at t . This method will cause only small errors at the inter-

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mediate time steps, since the wind disturbances are of large scale, and also varied very slowly.

For the fine-mesh computation, we also used quadratic interpolation with respect to the i, j coordinates. In order to obtain fine-mesh values in an area ABCD, we used the wind-stress values at six rough-mesh points A, B, C, D, P and Q (cf. Figure 17).

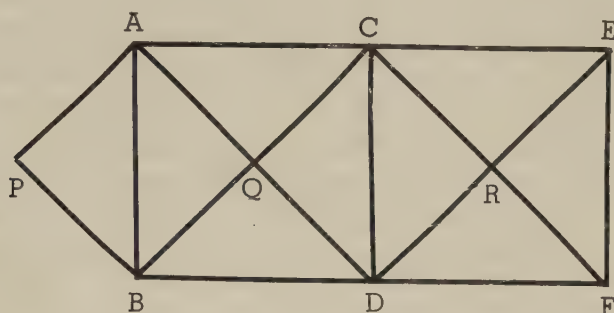


Figure 17. Explanations of the interpolation scheme.

(iii) Elevation at the seaward edge

Linear interpolation is adopted with respect to time and space. In this case, we use h -values only for even j -values, since these are nearly independent of the odd- j h -values in our case, and the mixed use of these two sets of values would cause computational errors.

Appendix II

Energy balance on the continental shelf in the case of Hurricane Carla

For the purpose of fine-mesh computations on the northwestern continental shelf, assumption was made that the elevation at the seaward edge of the shelf is given by the results of rough-mesh computations for the whole Gulf. This assumption requires that the energy flow across the continental slope should be small and effected little by the phenomena on the shelf.

In order to examine requirements, we computed the energy balance in this area. The results of wide-mesh computations were used for this purpose. The computation shows that the rate of energy generated on the northwestern shelf ($D \leq 100$ fm) increases with the approach of the hurricane center, reaches a maximum value 8.8×10^{18} erg/sec at 09Z September 11, and then rapidly de-

creases with time, while the energy flow is negligibly small until 18Z September 10. Slight inflow then appears and continues for about 12 hours. The maximum inflow was 1.1×10^{18} erg/sec, which is only one eighth of the rate of energy generated in this area. In the later stages, the energy flow turns to seawards, but the magnitude is very small except for the last few hours when the hurricane already reached the coast and storm surges become smaller.

These results indicate that the energy flow across the northwestern continental shelf is actually very small compared with the rate of work done by the external forces in this area, if we disregard the final few hours during which the storm surges are already diminishing. The direction of the energy flow across the continental shelf was also shoreward until the maximum of the rate of work by wind and pressure forces appears. This also will show us that the phenomena outside the shelf influence only to small extent the estimate of highest surges.

It will be of interest to note that 12-hour period oscillations are present in these time variations of energy.

Appendix III

Additive numerical experiments by changing the values of the bottom and surface stresses

As stated in Section 3 of this report, we adopted the following values of the bottom and surface stress factors:

$$\gamma_b^2 = 2.6 \times 10^{-3},$$

$$h_{s,b} \equiv k(1 + 5\gamma_b) = 3.1 \times 10^{-6} \text{ (c.g.s.)}.$$

However, these values are by no means well-established ones, but we have ample possibilities of getting different values or functional forms when our knowledge is improved.

At this point of view, we also tried some other coefficients. Some of our results are shown in Figure 18. In this figure, two full-lined curves show variations of the maximum computed surges at Galveston and Pass Cavallo for above $k_{s,b}$ -value and four different values of the bottom friction

$$\gamma_b^2 = 0, 1.0, 2.6 \text{ and } 3.9 (\times 10^{-3}).$$

These curves show that variations of the maximum computed surges are not so sensitive for the variations of the bottom friction for a wider range of γ_b^2 ($1 \sim 4 \times 10^{-3}$), but neglecting friction entirely gives significantly different results from a value greater than 1×10^{-3} .

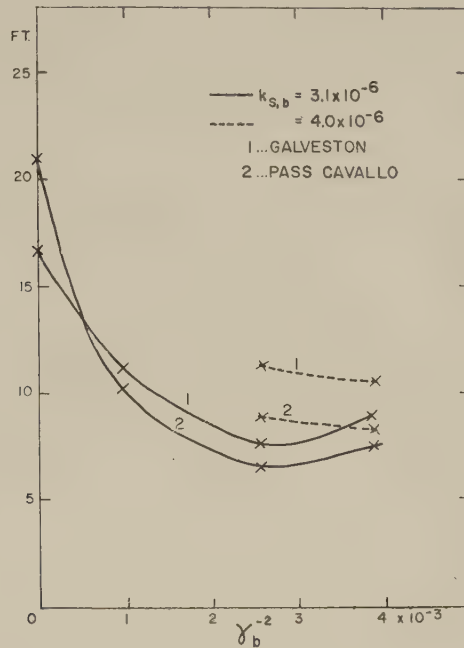


Figure 18. Variations of the maximum computed surges at Galveston and Pass Cavallo by changing the surface and bottom stresses.

In contrast, current speeds varied considerably as the bottom friction was changed, The maximum computed speeds⁶ are 6.7, 4.2 and 3.2 knots against $\gamma_b^2 = 1.0, 2.6$, and 3.9×10^{-3} , respectively.

The bottled two curves indicate variations of maximum surges for a different value of $k_{s,b}$. The relative increments of the maximum surges are roughly comparable to those in the $k_{s,b}$ -values.

It will be concluded from these experiments that maximum computed surges will change little by the variations of the bottom friction when γ_{bs} is the order of our estimation.

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6. 50 to 60 nautical miles south of Pass Cavallo.

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The Observation of the Internal Waves

By

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Abstract: Observations of internal waves were carried out on board R.V. Shumpū Maru from 1960 to 1964. The instrument used was mainly bathythermograph. The observations were made almost in the cold water region or near the coast south of Japan, where the thermocline was relatively shallow. The characteristic feature of the internal wave was that it had the periods of tidal or inertia motion. But their values resembled so closely with each other that we could not divide them into two parts clearly.

The surface current measured with GEK also varied, and its range was about 0.8 knot in all cases.

The heat content of the water column from the sea-surface to the depth of 250 m varied with the internal wave about 20,000 ly.

Steric sea level anomalies were also changed and hence geostrophic current was changed with time (about 50 cm/sec.).

1. Introduction

It is well known that there are various kinds of waves in the stratified ocean not only at the surface but also in the interior of the sea. The latter one is called the internal wave and is known to be excited by the tide-generating force, oceanic current, wind or atmospheric pressure disturbance. This phenomenon has already been investigated by many authors, i. e., FJOLDSTAD (1936), HAURWITZ (1950), MUNK (1945), DEFANT (1950, 1952), REID (1956), YASUI (1958, 1960), KRAUSS (1957, 1958), LAFOND (1962), RATRAY (1957, 1960), SUMMER *et al.* (1963) and so on.

This kind of wave progresses very slowly and has a large amplitude. In spite of these many studies, the character of the internal wave has not yet been completely known because of the scarcity of data.

We tried to observe the internal wave by several methods and studied its character. The speed of it was as much the same as the current or of the drifting vessels, so the observation of it was carried out very carefully. We hoped to anchor the survey vessel, but unfortunately our vessel had no equipment to anchor

herself in the depth more than 20 m, so we had to observe the phenomenon from the drifting or sailing vessel. In general, we may consider that the fluctuation of the isotherms shows the fluctuation of the isopycnal surface, that is to say, an internal wave. Then we may use BT record to observe the internal wave. As the surface current should be changed by the internal wave, we can observe it with GEK. In the following sections, a description will be made of the results of observations.

2. Observations

Figure 1 shows the observation stations from 1960 to 1964. The results of the investigation for the stations A and U in this figure were already reported (KONAGA, 1961). Then we will here report the results from 1961 to 1964.

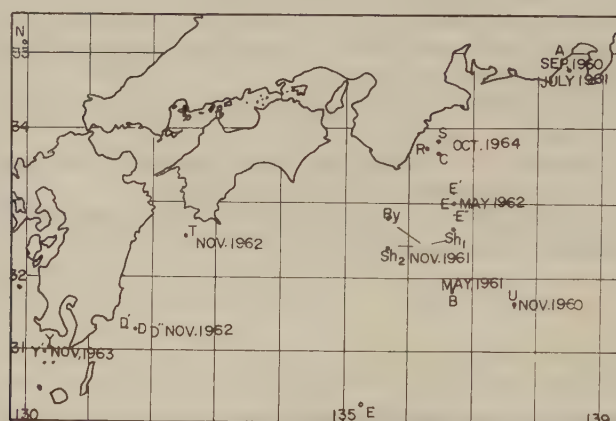


Fig. 1. Locations of the observation stations of internal waves.

a) Station B (31° - 50° N, 136° - 40° E), 23-24 May 1961

BT was cast in every 4 hours. This station is located in the cold water region near the northern edge of Kuroshio off Kumanonada (Fig. 2). At this station, the half pendulum day is about 23.1 hours and the periods of M_1 and M_2 tides are about 24.8 and 12.4 hours, respectively. The surface current varied clockwise, its maximum value being 2.1 kt and minimum value, 0.6 kt (Fig. 3). The depth of isotherms varied within the range of about 30 meters, with mainly diurnal period. A period of 8 hours appeared and the wave had a pretty large amplitude. The appearance of this wave was probably due to a small cyclone passing near the station in the midnight of the 23 rd.

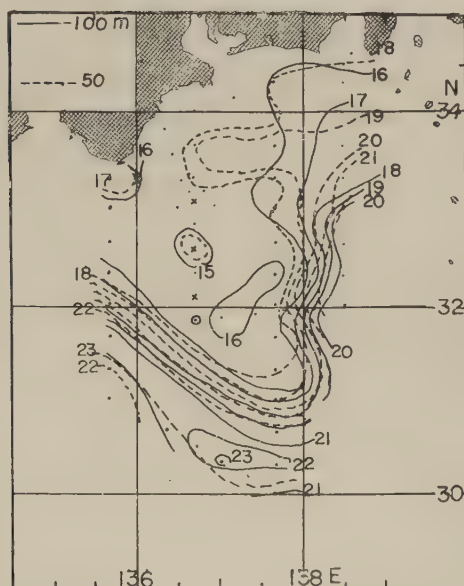


Fig. 2. Horizontal distribution of water temperature in May 1961 (full line: 100 m-layer; broken line: 50 m-layer).

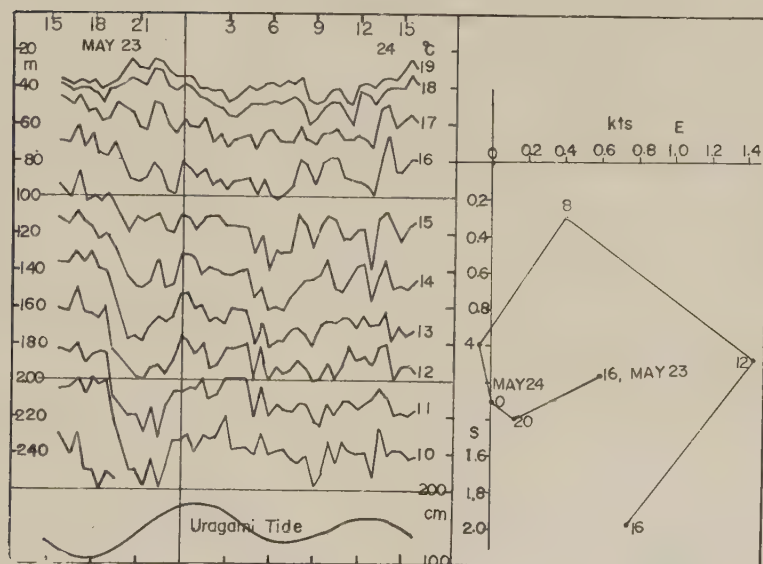


Fig. 3. Fluctuations of the isotherms (left) and surface current (right) at the station B.

b) Station A (in Suruga Bay), 16-17 July 1961

BT was cast in every one hour and GEK measurement was made in every 4 hours. In September 1960, at the same station we observed an internal wave of about semidiurnal and of large amplitude, but now we could not observe any distinct wave form in the fluctuation of isotherms (Fig. 4), for the instrument

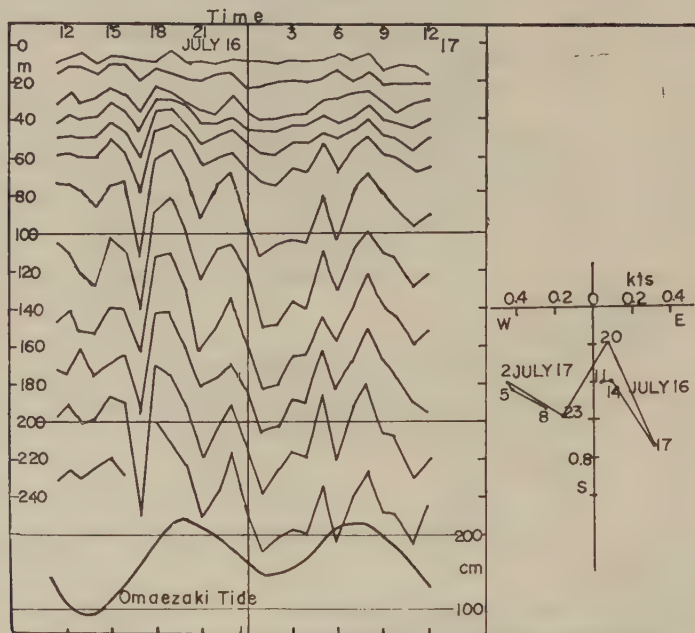


Fig. 4. Fluctuations of the isotherms (left) and surface current (right) at the station A.

was not so good for the present purpose for this season, namely the seasonal thermocline appeared near the surface (within 10 m from the sea surface) and the internal wave was disturbed by a small error in the surface water temperature. Though the short-period fluctuation and the error disturbed the wave form, there existed semidiurnal fluctuation in isotherms.

The surface current changed irregularly, because the record of GEK was affected by the tidal current and the change of the earth electric current (Cox, *et al.*, 1964).

c) Stations Ry (R. V. Ryofū Maru) and Sh₁ (R. V. Shumpū Maru) located at 32°-50'N, 135°-40'E and 32°-40'N, 136°-40'E, respectively, 8-10 November 1961

In order to study the character of the internal wave, we made an observation

on board two vessels about 50 nautical miles apart from each other. Both stations were located at the points about 20 miles distant to the north of Kuroshio and not affected by it in such a short period (Fig. 5). Shumpū Maru alone measured the surface current with GEK in every 3 hours. BT was cast in every one hour and serial observation was made in every 3 hours. Before this observation was carried out, a cross observation near the station Sh_1 was made. The fluctuations of the isotherms were depicted in Figures 6 and 7. The internal waves at each station did not show an explicit interrelation. Their amplitudes were smaller than we expected. As regards the less close relationship between them, we might think probably that both ships were too far apart from each other to observe the internal wave and oceanographic

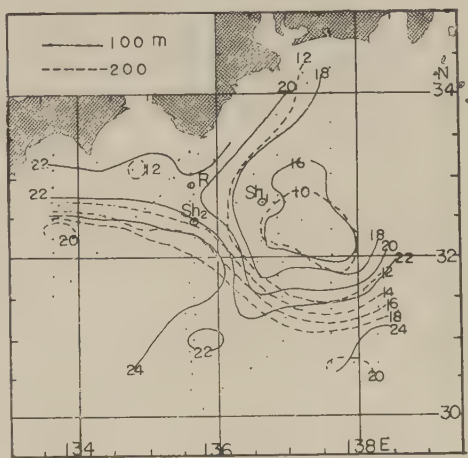


Fig. 5. Horizontal distribution of water temperature in Nov. 1961 (full line: 100 m; broken line: 200 m).

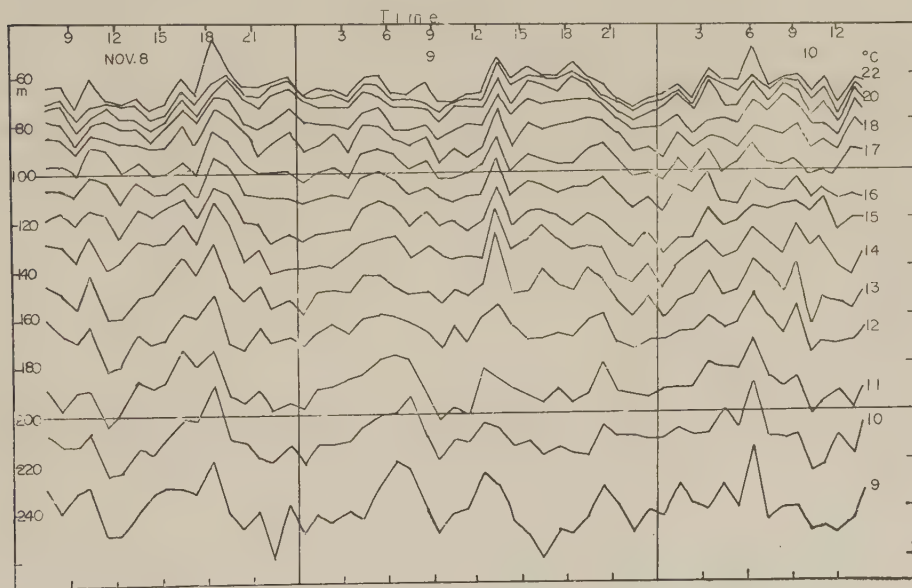


Fig. 6. Fluctuations of the isotherms at station Sh_1 .

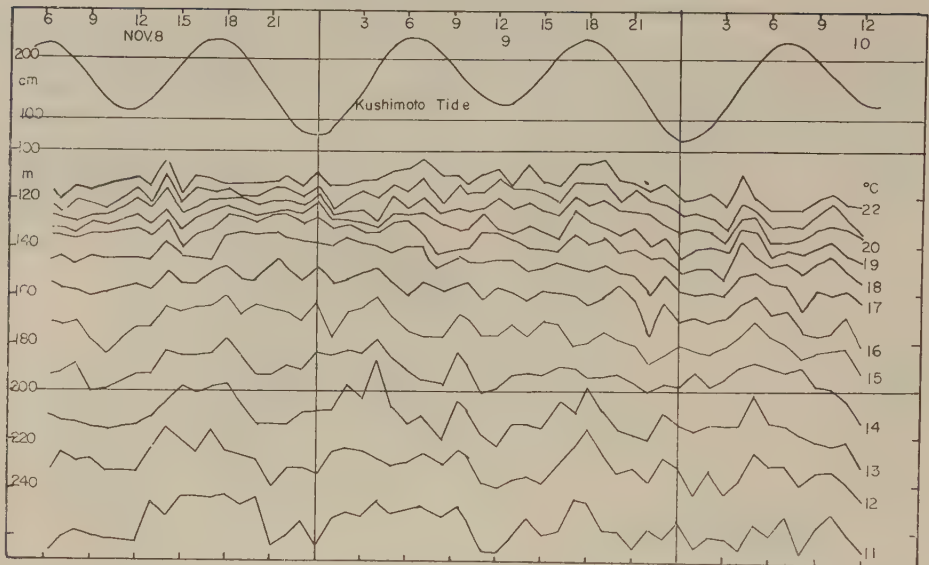


Fig. 7. Fluctuations of the isotherms at station Ry and surface tide at Kushimoto.

condition was not coincident with each other, that is, Sh_1 was located in the cold water region, while Ry, in coastal water region, near Kii Peninsula, and so the surface mixing-layer-depth was deeper by about 30 m at the station Ry than Sh_1 . Owing to these differences of oceanographic conditions the wave may not have progressed in the same form but changed its speed and form, when it crossed the Kuroshio counter current flowing southward along the meridian of $136^\circ E$ in this latitude. Thus, our trial to determine the three elements of the internal wave, that is, the wave height, the direction and speed of progress and the period, was not successful, but it has been suggested by this trial that for this purpose the observations of the internal wave should be taken by many vessels situated 10 or 15 miles or so apart from each other and in the sea of the same oceanographic condition.

The main result of the present observation was as follows: At the station Sh_1 , internal waves with the periods of two days and 16 hours existed near the surface though their amplitudes were small. In deeper layer ($10^\circ C$) there existed internal waves with nearly diurnal and semidiurnal periods. At the station Ry, there existed internal waves having the periods of one day and two days near the surface, and in deeper layer ($12^\circ C$), the internal waves of semidiurnal period appeared, though their amplitudes were small, too. At both stations, the two-

day-period was dominant, but this wave had no physical meaning. Probably the mathematical process yielded these results. The surface current at station Sh_1 rotated clockwise with the radius of about 0.5 miles (Fig. 9).

d) Station Sh_2 (32° - $25'$ N, 135° - $40'$ E), 15-16 November 1961

This station was located at the northern edge of the Kuroshio and so the lateral gradient of the current and the water temperature were pretty large. So the vertical oscillation of the isotherm appearing in the BT records could not be decided to be an internal wave, because of the uncertainty of the location of the vessel and the variation of the oceanographic condition with short period. When the location of the vessel was shifted about 5 miles, the water temperature at the depths of 100 m and 200 m changed about 1.0 and 1.5°C , respectively, and surface current changed about 0.5 kt in mean current speed. The result is shown in Fig. 8. If we suppose that these variations were all consisted of internal waves, its

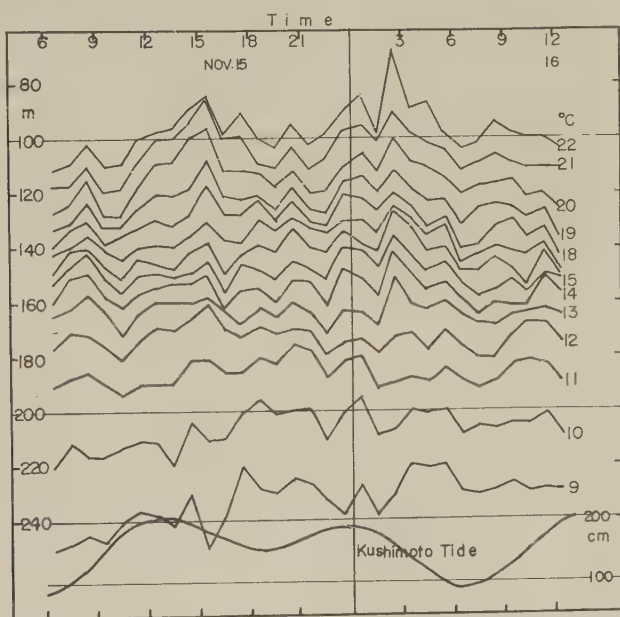


Fig. 8. Fluctuations of the isotherms at station Sh_2 .

amplitudes were very large and the semidiurnal wave was predominant. In Fig. 9, the surface current is shown; the upper: the case (c), lower: the case (d). In the one the current varied clockwise with a period about the half pen-

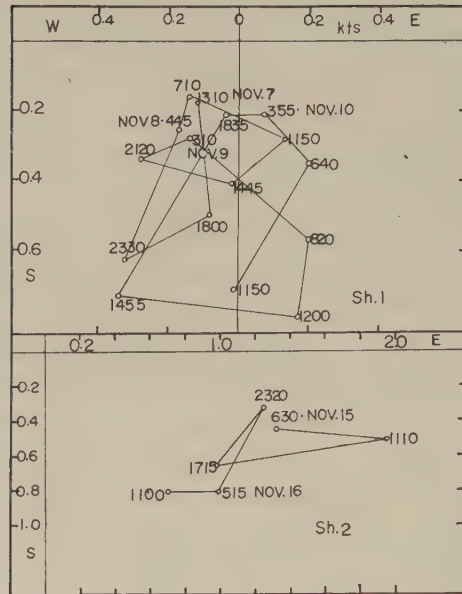


Fig. 9. Fluctuations of the surface current at stations Sh₁ (upper) and Sh₂ (lower).

dulum day, and in the other, it seems like to vary clockwise but not distinct. In the case (c) the surface ordinary tide was spring, and in the case (d), neap, so the results suggested that the fluctuations of the isotherms consisted not always of the internal tide, but of inertia wave, or merely a lateral uncertainty of the position of the vessel.

e) Station E (32° - $40'N$, 136° - $40'E$), E' (about 8 miles north of St. E) and E'' (about 8 miles south of St.E), 30-31 May 1962

BT cast was made in every 2 hours at the centered position (station E) and the others (station E' and E'') in every 4 hours. GEK observation was carried out at every station. The stations were all in the cold water region and lateral gradient of the water temperature and the surface current were very small (Fig. 10). The purpose of this observation was to study the direction of propagation of the internal wave. If the internal wave consisted mainly of internal tide, as we assumed that it proceeded to east-west direction, it was probable that the phase of fluctuation was almost coincident with each other in every station. But if it consisted of inertia wave, we assume that it would proceed to the current

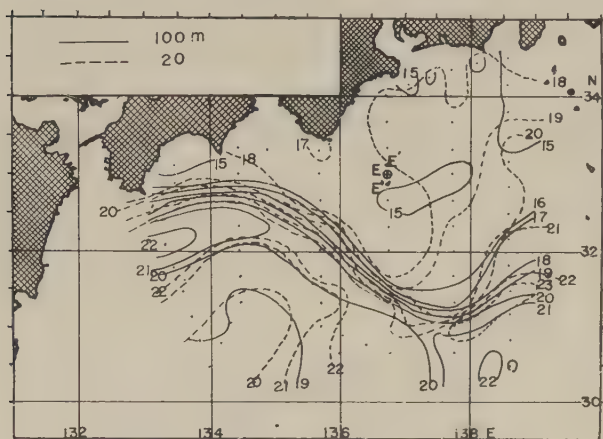


Fig. 10. Horizontal distribution of water temperature near station E, in May 1962 (full line: 100 m; broken line: 20 m).

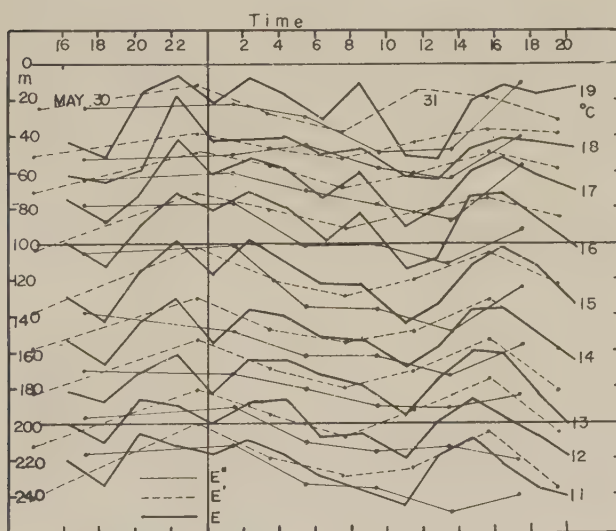


Fig. 11. Fluctuations of the isotherms (thick line: station E; thin line: station E'; broken line: station E'').

direction or randomly, the phase was not coincident. The results were as follows: The periods of them were a half day and a day. Especially, the diurnal wave was predominant at all stations. From the scarcity of data, the relation of the phases was not appreciated. During this observation, in the day time in 31 May, the surface water temperature rose abruptly about 4°C . At this time the current

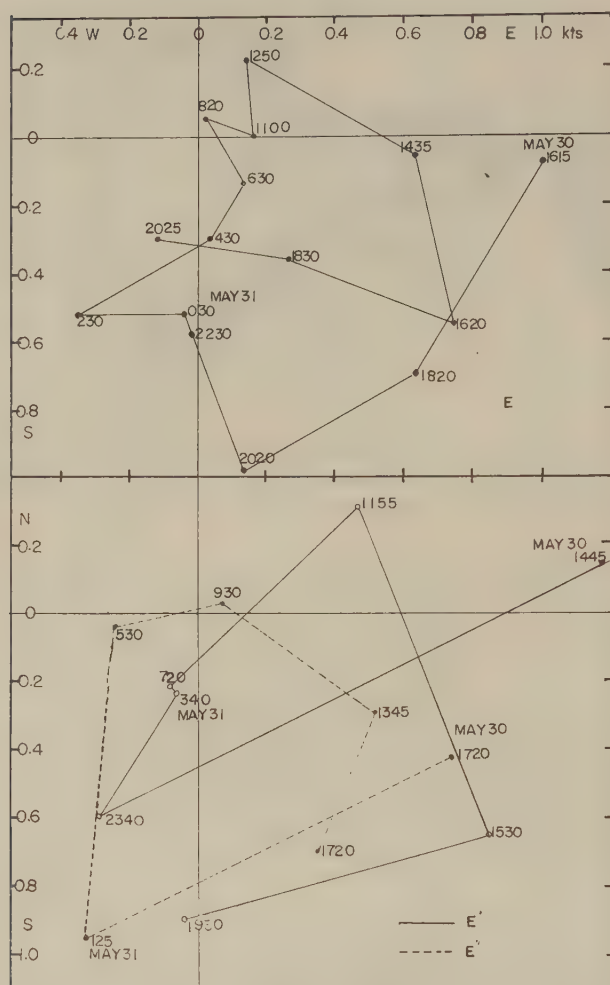


Fig. 12. Fluctuations of surface current (upper: station E; lower: full line: station E'; broken line: station E'').

was very weak and the internal wave made convergence of the surface water and the dynamical stability of the surface water was great. The strong solar radiation could rise the water temperature in nearly static state. This result had already been reported (KONAGA, 1962).

f) Station D ($31^{\circ}-17'N$, $131^{\circ}-43'.5 E$), D' ($31^{\circ}-18'.5 N$, $131^{\circ}-36'E$) and D'' ($31^{\circ}-15'.5 N$, $131^{\circ}-52'E$), 6-7 November 1962

By the same method as in the case (e), now we select the station near the

Kuroshio. The purpose of this observation was to study the inertia wave and the short period fluctuation of the Kuroshio near the coast. The distributions of the water temperature at 50 m and 100 m depth are shown in Fig. 13. The lateral gradient of the water temperature was very large below 100 m depth. The results were as follows: The short period fluctuation of the Kuroshio was shown by the gradual fall of the water temperature in all stations (Fig. 14). The fluc-

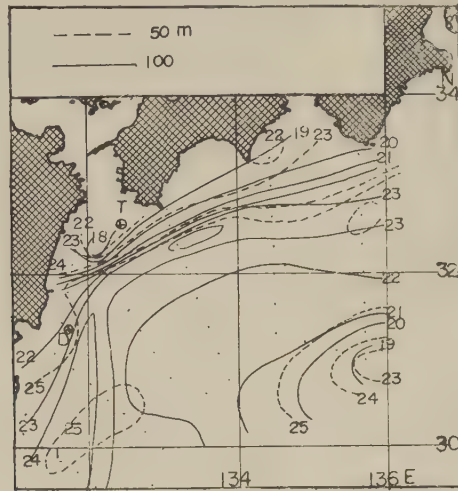


Fig. 13. Horizontal distribution of water temperature in Nov. 1962 (full line: 100 m; broken line: 50 m).

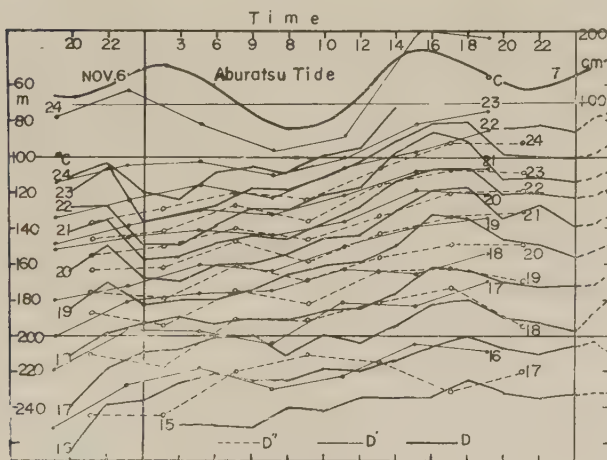


Fig. 14. Fluctuations of the isotherms at station D (thick line), station D' (chain line) and station D'' (broken line).

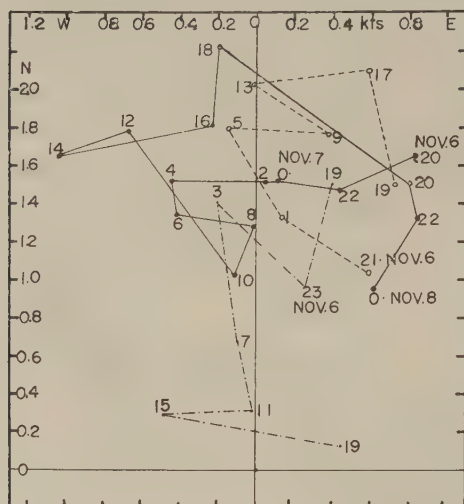


Fig. 15. Fluctuations of surface current (full line: station D; chain line: station D'; broken line: station D'').

tuation of the isotherms showed about 18 hours period. The surface current changed with clockwise rotation of about 23-hours' period (Fig. 15). These results suggest that the fluctuation of the isotherms was a combination of the internal tide and the inertia wave.

g) *Drift observation in the Kuroshio from the station D (i: 7-9, ii: 9-10 and iii: 10-11 November 1962)*

(i) D₁: At the first time the sea condition was pretty calm and wind was weak. The purpose of this observation was to find the internal tide near the coast, progressing away off shore (or toward coast). If the inertia wave was produced by the current, it probably proceeded with the current with almost the same velocity. And the observed fluctuation of the isotherm should mainly be due to the internal tide. The wake of the vessel was shown in the right part of Fig. 16. In Fig. 17, the water temperature was slightly fall but it showed that 24°C-isotherm varied in phase with the tide at Aburatsubo. The other isotherms were not in phase and had rather small amplitudes.

(ii) D₂: In Fig. 18, the wake of the drift of the vessel (about 8 hours, repeated) was shown. During the observation, wind of about 7 m/sec. blew. The result was shown in Fig. 19. Though the variation of the range of water temperature was very large, its periodicity was obscure, probably because of the short

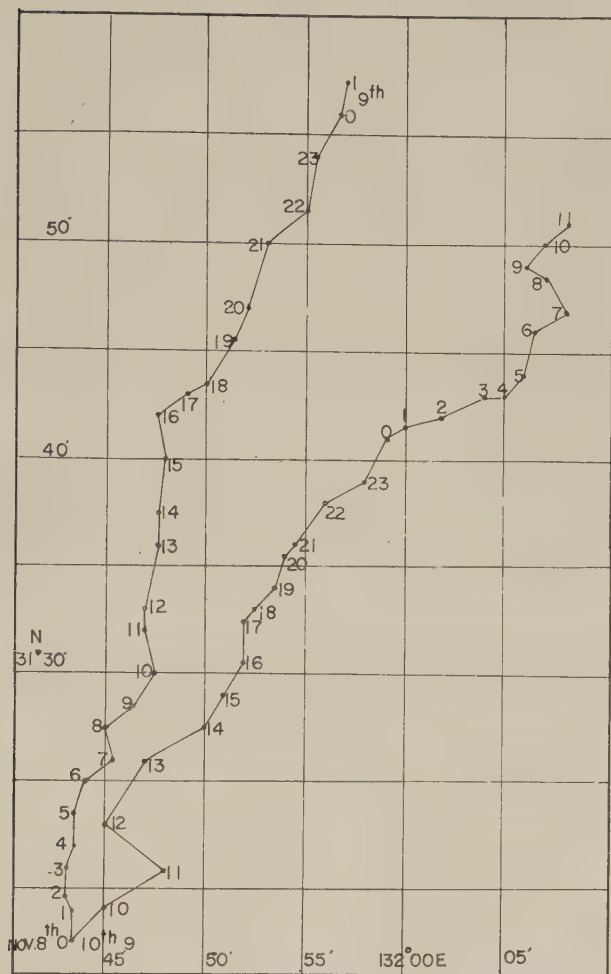


Fig. 16. Wakes of the drifting vessel from station D

period fluctuation of the Kuroshio and the discontinuity of the inertia wave.

(iii) D_3 : The right-hand side of Fig. 19 showed the drift of the vessel from Nov. 10 to 11. During this observation the wind blew at about 9 m/sec. and the vessel was drifted with wind and shifted from the Kuroshio current. From Fig. 20 we can see that the water temperature rose slightly and periodic motion was very weak. But below the 200 m-depth the mean temperature did not vary and a periodic motion appeared, whose period was about 11 and 17 hours. It was probable that the wind generated another inertia wave. In cases (ii) and (iii) the

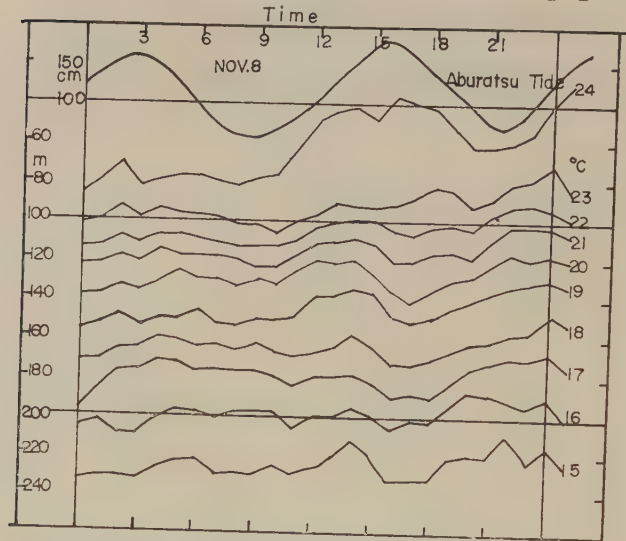


Fig. 17. Fluctuations of the isotherms and surface tide at Aburatsu in the case D_1

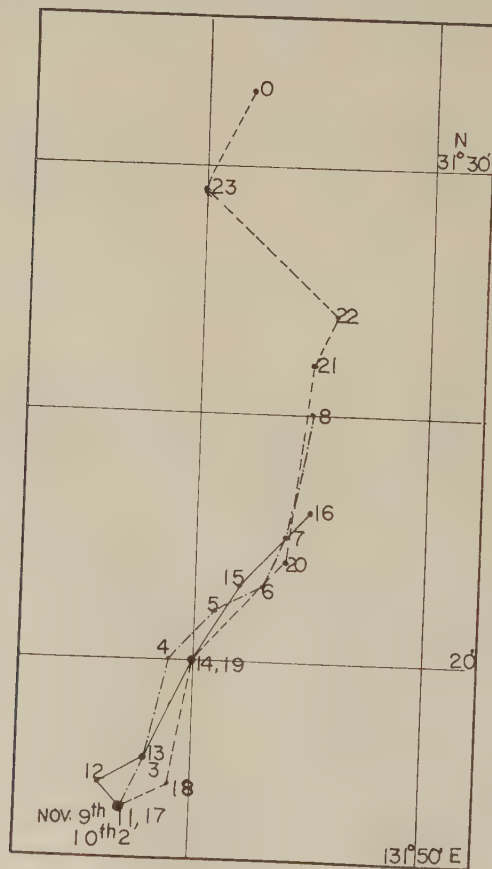


Fig. 18. Wakes of the drifting vessel from station D in the case D_2 .

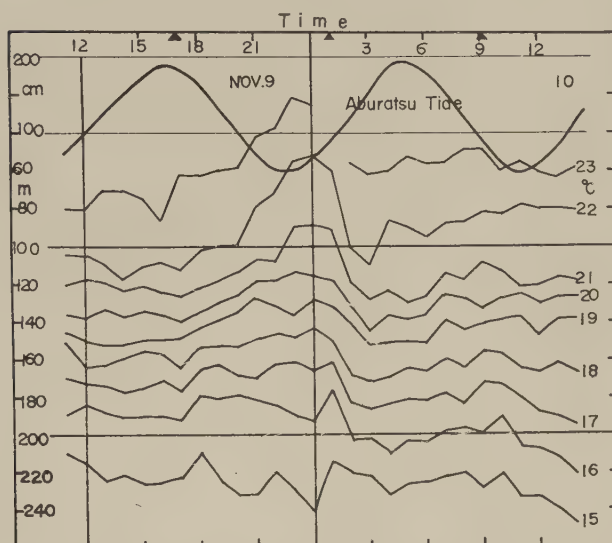


Fig. 19. Fluctuations of the isotherms in the case D_2 and tide at Aburatsu.

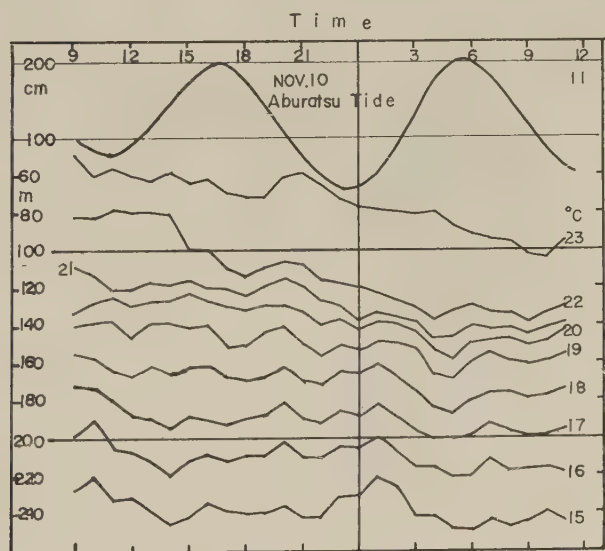


Fig. 20. Fluctuations of the isotherms in the case D_3 and tide at Aburatsu.

fluctuation of the isotherms were not in phase with surface ordinary tide, as shown in those figures. Hence we can not conclude that the internal tide proceeded in east-west direction across the Kuroshio.

h) Station T (32° - $32'N$, 132° - $30'E$), 18-20 November 1962

BT was cast in every one hour, the serial observation to the depth of 400 m was carried out in every 4 hours and GEK observation also in every 4 hours. The vessel was drifted and shifted to correct the position in every 4 hours. The drift velocity and the measured velocity were compared with each other by the author (KONAGA, 1964). In spite of the coastal region the internal wave had only a small amplitude. The periods were a half day and shorter. To be apart 4-5 miles from the original position, the internal wave would be disturbed.

i) Station Y (31° - $02'N$, 130° - $22'E$) and Y' (30° - $58'N$, 130° - $17'.7 E$), 16-19 November 1963

In the first 24 hours at the station Y and Y', BT and GEK observations were carried out in every 2 hours respectively and in every one hour at the mid-point of both stations (Y''), GEK was alone observed. In next 48 hours the the drift observation (BT, in every one hour; serial and GEK observations, in every 4 hours) was carried out. The purpose of this observation was to find the internal wave on the oceanic ridge. The oceanic condition is shown in Fig. 21. The lateral

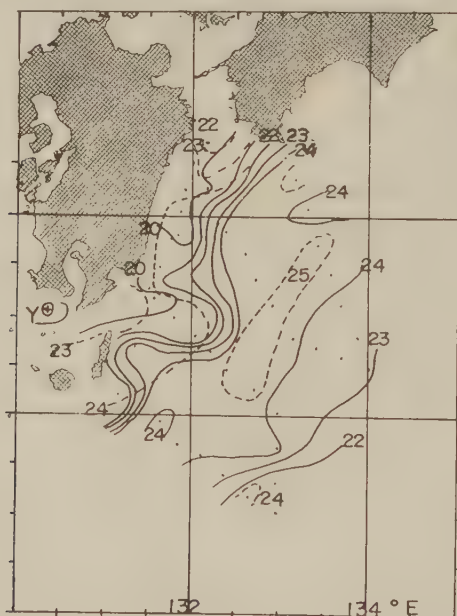


Fig. 21. Horizontal distributions of water temperature in Nov. 1963 (full line: 100 m; broken line: 50 m).

gradient of the water temperature was weak, but the coastal and off shore water were contacted near this station. So in the surface layer the water temperature

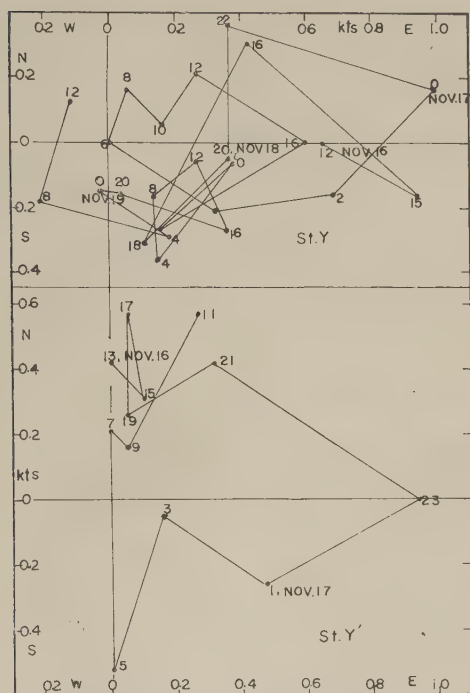


Fig. 22. Fluctuations of surface current at stations Y and Y'.

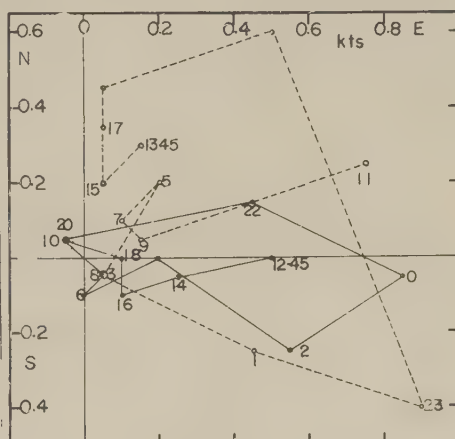


Fig. 23. Fluctuations of surface current at the mid position between Y and Y'.

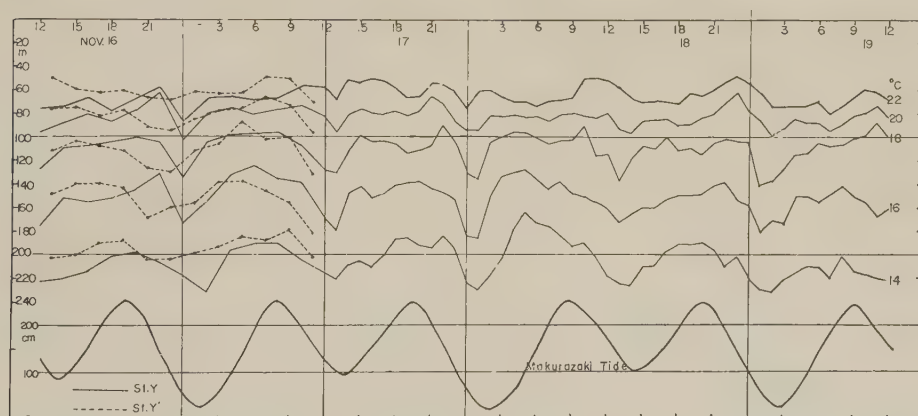


Fig. 24. Fluctuations of the isotherms (full line: station Y; broken line: station Y') and tide at Makurazaki.

changed abruptly about 1.0°C but not affected to depth below 50 m. The results were as follows (Figures 22, 23 and 24): The current varied clockwise at each station and their periods were 10 to 26 hours. Probably ordinary tidal current and inertia rotation appeared on these records. The fluctuation of the isotherms seems to show the tidal period, approximately in phase with ordinary tide at Makurazaki. From the periodgram analysis the predominant periods were about 11.4 and 23.7 hours, which did not coincide with neither tide nor inertia period. The difference of phases between the stations Y and Y' was about 1.5 hours. If the wave progressed in east-west direction, the wave progressed toward east at about 2.5 kt or toward west at 0.4 kt. The former seems to be more reasonable.

j) RSC cooperative observation off Kumanonada, R ($33^{\circ}-46'.1\text{ N}$, $136^{\circ}-18'.1\text{ E}$), S ($33^{\circ}-50'.4\text{ N}$, $136^{\circ}-27'.0\text{ E}$), and C ($33^{\circ}-41'.0\text{ N}$, $136^{\circ}-27'.0\text{ E}$), 20-25 October 1964

The three elements of the internal wave are the wave height, frequency or period and wave velocity, the direction and speed. The past observations which we had taken were mainly the height and period of the wave. The wave velocity would not be determined without more than three vessels.

Ryofū Maru (JMA), Shumpū Maru (KMO) and Chofū Maru (NMO) were gathered in Kumanonada to observe the internal wave near the east coast of Kii

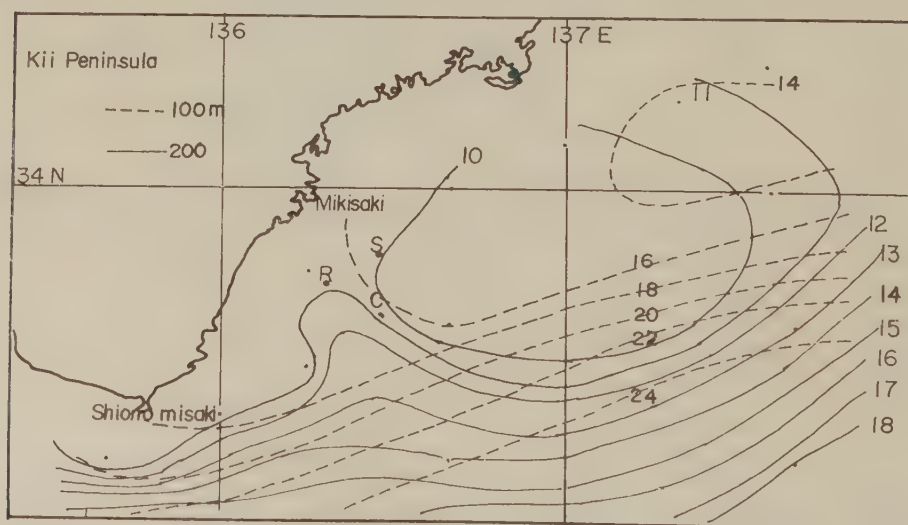


Fig. 25. Horizontal distributions of water temperature in Oct. 1964 (full line: 200 m; broken line: 100 m).

Peninsula. Ryofū Maru (St. R) was anchored at the station where the depth of the sea was about 1500 m.; Shumpū Maru and Chofū Maru were drifted near the fixed position (Sts. S and C) within about 2 miles. The vessels were apart from each other by about 9 miles. The station locations of R, S and C and the horizontal distributions of the water temperature were shown in Fig. 25. BT was cast in every one hour and serial observation to 1200 m was carried out in every 6 hours at each station, GEK observations were carried out in every 3 hours at the stations S and C. The lateral gradient of the water temperature was small, but lateral advection of coastal water sometimes appeared and chlorinity of the surface layer changed within a pretty wide range (about 0.3 ‰). The vertical distribution of chlorinity at each station is shown in Fig. 26. The fluctuation of the isotherms

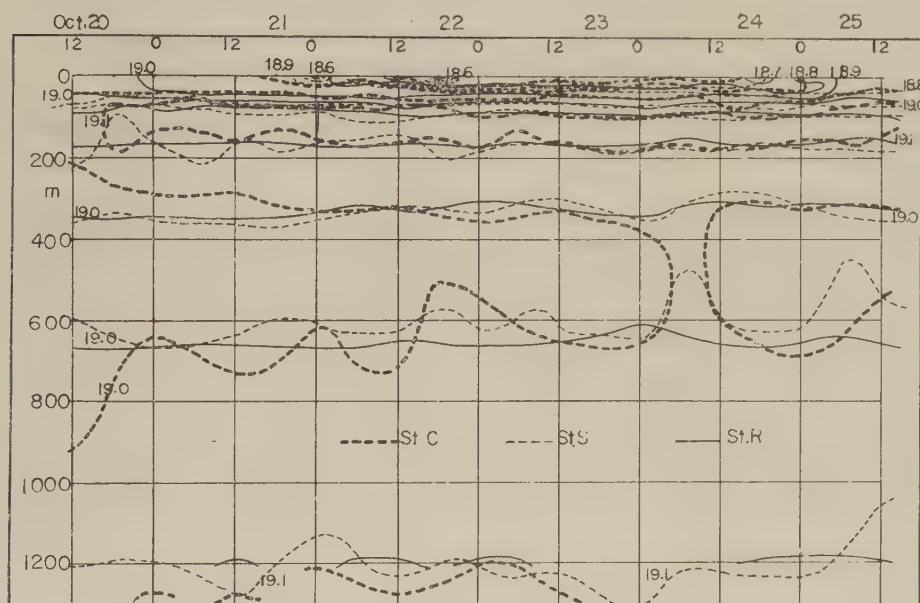


Fig. 26. Isopleth of chlorinity (full line: station R; thick broken line: station C; thin broken line: station S).

at each station is also given. BT records are shown in Figs. 27 (R), 28 (S) and 29 (C). Three-hourly moving averages were calculated and are shown in Fig. 30. At the station R, the fluctuation of the isotherm of 12°C was nearly periodic and its period was about 11.7 hours. At the station C the fluctuation of the 12°C-isotherm was nearly of the same form as above. But at the station S, such

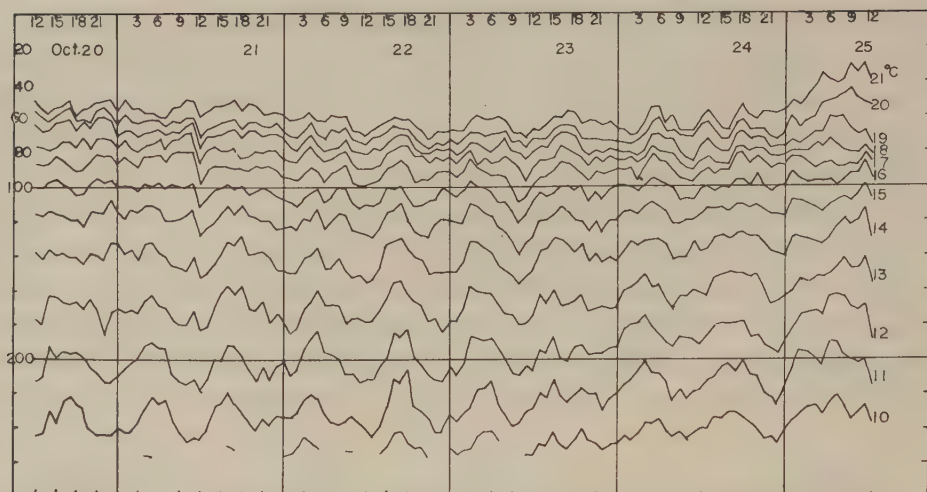


Fig. 27. Fluctuations of the isotherms at station R.

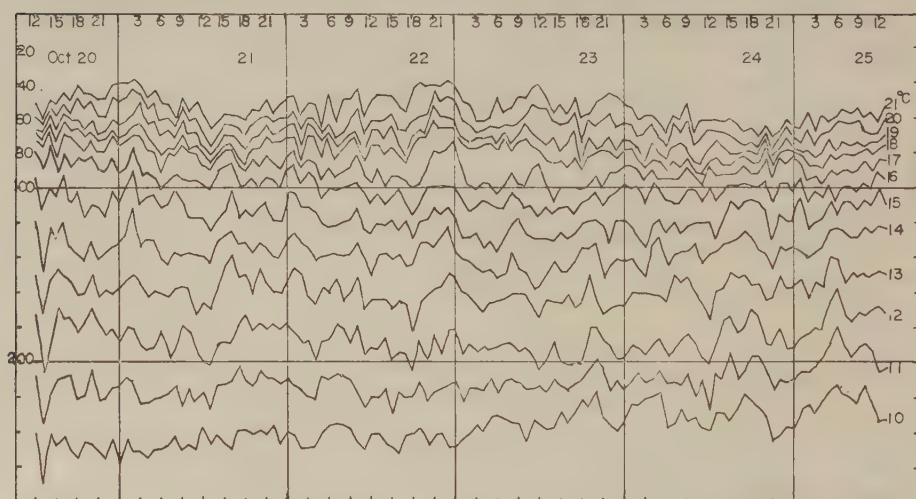


Fig. 28. Fluctuations of the isotherms at station S.

semidiurnal fluctuation disappeared without the center of these observations. Especially in the first 48 hours, there was no fluctuation having semidiurnal period. In the last 36 hours, the outburst of the northerly monsoon disturbed the oceanographic condition. And the water temperature fell except near the surface at station C. From the meteorological data in this period, it was known that at station S the strongest wind was recorded. The drift of the vessel became larger

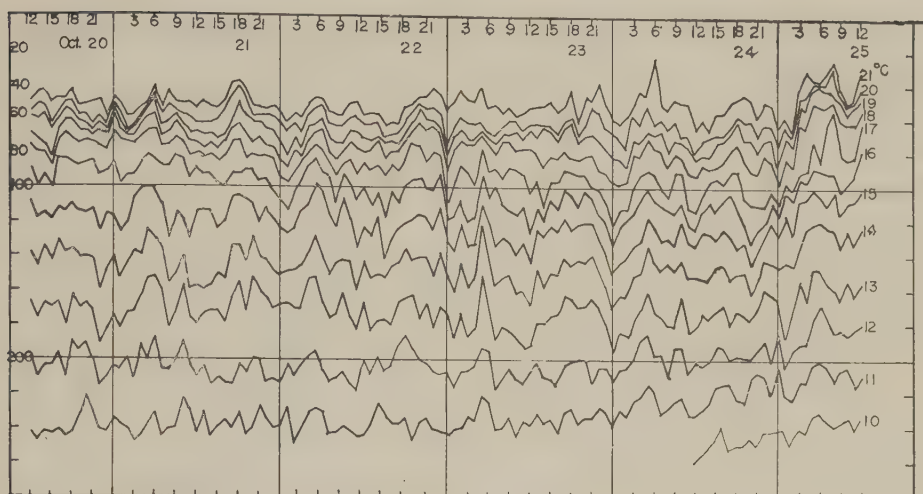


Fig. 29. Fluctuations of the isotherms at station C.

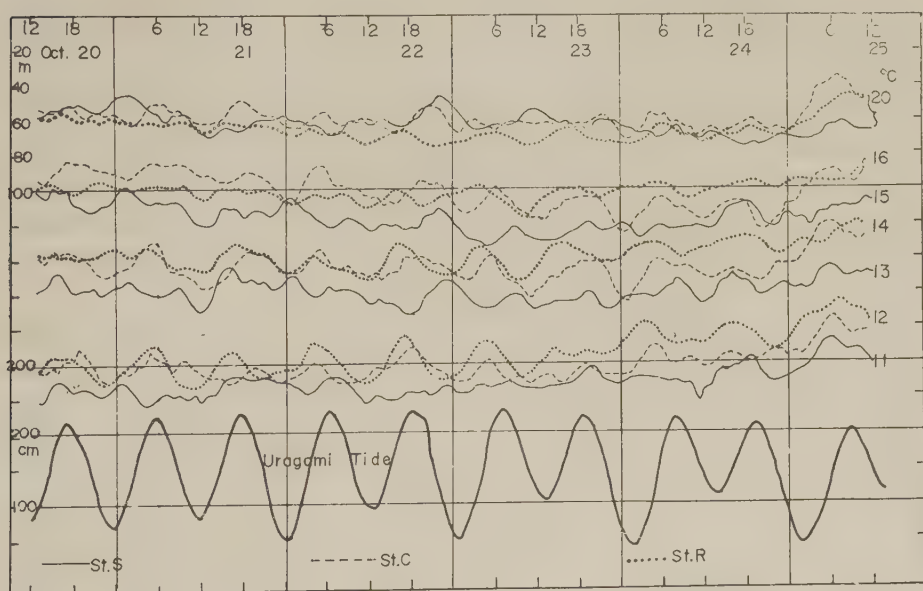


Fig. 30. Three-hourly moving average of the depth of the isotherms and tide at Uragami (full line: station S; broken line: station C; dotted line: station R).

and so the observed form of the internal wave was disturbed strongly. Hence only in the calm condition of the middle 48 hours, the useful data to calculate the wave speed or wave direction were provided. The periodgram analysis, coherency

and harmonic analysis were carried out. The predominant periods for each station were as follows: R: 11.7; S: 21.8; and C: 24.0 hours, respectively. For the wave of 11.7 hours-period, the phase lag at each station was as follows: R: 5.9; S: 0; and C: 7.0 hours, respectively. R and S differed about a half period. So S might be thought to be 11.7 hours. These two cases were considered in order to find the direction of propagation of the internal wave. To the former case, the internal wave proceeded to SSW at the velocity of about 1.3 kt and the others to NE at 1.6 kt. If we considered the non-rotating earth and two-layered ocean with density difference of about 0.02, the speed of the internal wave was about 1.4 kt. It takes about 6 hours between two stations. From this point of view, we neglect the case more than 8 hours and less 5 hours. Both cases of the above results fulfilled this condition. As the effect of the rotation of earth would increase the speed of the internal wave, we may say the wave proceed toward NE at 1.6 kt. If we considered the perfect data, though the periodicity and amplitude were obscure at station S, it progressed toward WSW at about 1.4 k. In general, supposing the internal tide, it would have progressed in east-west direction, and supposing inertia wave caused by the Kuroshio, it would have progressed to N or NE, and that caused by the meteorological disturbance, it would probably proceed in east-west direction, according as the disturbance moved toward east.

The observed period was neither tidal nor inertia, but the combination of lunar semidiurnal and a quarter pendulum day, which have no physical meaning, but has nearly equal period. More precise observation or analysis must be taken in order to decide the character of the internal wave.

The current variations were shown in Fig. 31; left: St. S and right: St. C. At station S, they varied in clockwise sense but its periodicity was not clear. They were divided into three groups by its velocity; each group was consisted of about 12 observations (about 36 hours). In all cases the mean current flowed toward SW and its speed was 0.5, 0.9 and 0.1 kt, respectively. At station C they varied rather randomly compared with station S. They were also divided into three groups. The first of them was about 25 hours, SSE 0.5 kt, next, about 70 hours, SW 0.8 kt, and the last about 25 hours, W 1.2 kt. So these variations were based on not only the internal wave but also the change of the general oceanographic condition.

We considered that the observations of the internal wave in the open ocean should be carried out with more than three vessels placed about 5 miles apart. To fix the positions of the vessels very exactly, we should anchor the vessels. And

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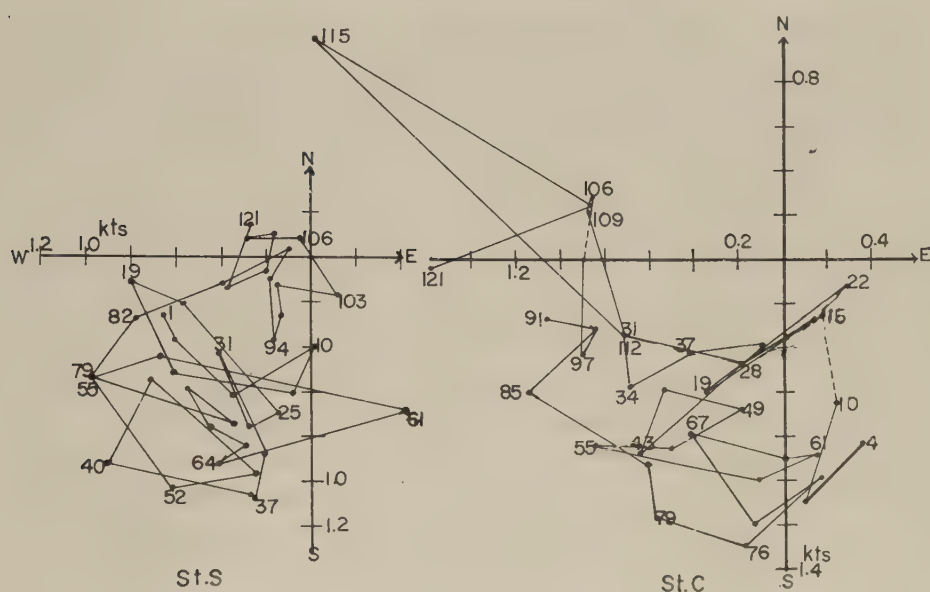


Fig. 31. Fluctuations of surface current at station S (left) and station C (right).

we could get more excellent data. In Table 1, we listed the predominant periods of the above cases obtained by the periodgram analysis. Since the data were too scanty to determine the period of the waves, the listed values are only approximate. And the difference of the wave form with depth makes a selection of large amplitude of waves.

Table 1. Predominant periods of the internal waves.

Case	Isotherm	Predominant period (hour)		Remarks
Sh ₁	10°C	12.5	24.5	ambiguous
Ry	12	12.0		
Sh ₂	20	12.5	24.5	
E	17	12.5	24.5	
D ₁	22	12		
D ₂	21		24	
D ₃	12	11	17	clear clear (at the central part, 11.5)
T	16	12.5		
Y	14	11.4	23.7	
R	12	11.7	23.4	
S	13	14.5	21.8	
C	12	9.8	24.0	

3. *The difference of the amplitude with depth*

It was seen from the figures that the fluctuations of isotherm did not have the same form for different isotherms. The wave form was not necessarily clear at the sharp thermocline, but in the deeper layer there existed fine wave form. If the wave were generated by the tide-generating force directly, the phase lag of the wave with depth might not appear. But if the cause of the internal wave were the tidal current flowing over the oceanic ridge or continental shelf and the inertia wave, the phase of the wave and the wave form would differ with depth. The amplitude, of course, has its maximum value at the layer, where the wave is generated, such as sea surface or the top of the oceanic ridge.

FJOLDSTAD explained the vertical distribution of the lateral current and the amplitude of internal wave from the density distribution. This result coincided well with actual distribution (MUNK, 1945). This suggests that the internal wave in actual ocean can not be explained by the hypothesis of the two layered ocean, but the continuous ocean. Moreover in the ocean there are many internal wave caused by many forces. It is probable that the free internal wave decayed very rapidly after proceeding for several wave lengths, but the forced internal wave was always produced here and there in the ocean. So we can recognize that even if many internal waves existed in the ocean, the explicit periodicity appeared in some depth. In other words, there was a depth where the internal wave disappeared by interference. In the surface layer, there probably was the inertia wave produced by the ocean current, wind and disturbance of the atmospheric pressure. Hence the internal wave near the sea surface was very complex compared with the deeper layer. In the cold water region the seasonal thermocline was shallower than 70 m-depth and the main thermocline was below the layer. BT record caught the internal waves both seasonal and main thermocline, but in the other region BT record would not be able to catch the main thermocline, at some place even the seasonal thermocline. So the former explains the internal tide or inertia wave, but the latter, probably only internal tide. This a very important problem to recognize the internal wave. Further investigations are necessary to explain this phenomenon.

4. *The variation of the thermal structure of the surface layer of the sea (Mean temperature near the sea surface)*

An internal wave existing at the sharp thermocline varied the depth of the surface mixing layer. If the ocean consists of an exact two-layer stratification,

the variation of the mean temperature $\left(T = \frac{1}{D} \int_0^D T dz\right)$ will be proportional to the amplitude of the wave, i. e., if h and H be the thickness of the upper and lower layers and T_1 and T_2 be the corresponding water temperatures, then

$$\tilde{T}_0 = \frac{1}{D} \int_0^D T dz = \frac{1}{D} \left\{ \int_0^h T_1 dz + \int_h^{h+H} T_2 dz \right\} = \frac{1}{D} (T_1 h + T_2 H), \text{ where } h + H = D.$$

If h increases to $h + \Delta h$, H decreases to $H - \Delta h$, and hence the mean temperature becomes

$$\tilde{T} = \frac{1}{D} \left\{ T_1 (h + \Delta h) + T_2 (H - \Delta h) \right\} = \tilde{T}_0 + \frac{\Delta h}{D} (T_1 - T_2)$$

Here Δh is thought to be the amplitude of an internal wave.

If the temperature decreases linearly with depth under the mixing layer and the amplitude and the phase of the internal wave be independent of the depth, then the above equation becomes as follows:

$$\begin{aligned} T &= T_0 - a(z - z_0) & z \geq z_0 \\ &= T_0 & z < z_0 \\ \tilde{T} &= \frac{1}{D} \left\{ T_0 z_0 + \int_{z_0}^D (T_0 - a(z - z_0)) dz \right\} = \frac{1}{D} \left\{ T_0 D - \frac{a}{2} (D - z_0)^2 \right\} \\ &= T_0 - \frac{a}{2D} (D - z_0)^2 \end{aligned} \quad (4.1)$$

If z and T vary with time, supposing that

$$z_0' = z_0 + b \cdot \cos \omega t,$$

equation (4.1) can be written in the following form:

$$\begin{aligned} \tilde{T} &= T_0 - \frac{a}{2D} (D - z_0 - b \cdot \cos \omega t)^2 \\ &= T_0 - \frac{a}{2D} \left\{ (D - z_0)^2 + b^2 \cdot \cos^2 \omega t - 2(D - z_0)b \cdot \cos \omega t \right\} \end{aligned} \quad (4.2)$$

As numerical values, we shall take

$$T = 21.5^\circ\text{C}, \quad D = 50 \text{ or } 250 \text{ m}, \quad a = 5/3000^\circ\text{C cm}^{-1} \text{ or } 10/20000^\circ\text{C cm}^{-1}$$

$$z_0 = 4000 \text{ cm}, \quad b = 1000 \text{ cm} \text{ and } \omega = \frac{\pi}{6}.$$

Substituting these values into equation (4.2),

i) $D = 50$

$$\begin{aligned} T &= 21.5 - \frac{5}{2 \times 3 \times 5 \times 10^6} (10^6 + 10^6 \cos^2 \omega t - 2 \cdot 10^6 \cos \omega t) \\ &= 21.5 & \text{when } t = 0 \\ &= 21.3 & t = 3 \\ &= 20.8 & t = 6, \end{aligned}$$

ii) $D=250$

$$T = 21.5 - \frac{1}{2 \times 2 \times 25 \times 10^7} (4.41 \times 10^8 + 10^6 \cos^2 \omega t - 4.2 \times 10^7 \cos \omega t)$$

$= 17.5$	when $t=0$
$= 17.1$	$t=3$
$= 16.7$	$t=6$

At $t=0$ and $t=6$, we considered the maximum and minimum values of the heat contents in the water column. The variations of the heat content in the water columns, of 1 cm^2 -section and of 50 m and 250 m depths, were about 3300 ly and 42000 ly respectively. But the phases of the internal waves were different for the different depths, and in spite of the fact that the observed internal waves had the wave height of about 20 m, they did not vary so much as might be inferred from the heat content computed above. Actually the differences of the heat content in the water column, of 1 cm^2 -section and of 50 m or 250 m depths, were about 2700 ly or 20000 ly in short term (in all the period of the case (j) in the former section, the maximum differences were 5500 ly or 37000 ly, because of the longer-period variation and the superposition of many internal waves). An example is shown in Fig. 32.

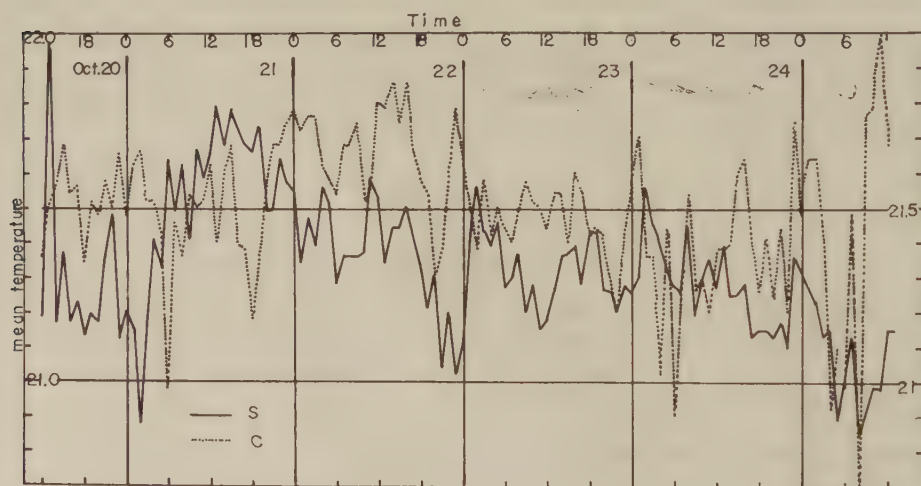


Fig. 32. Variations of mean temperature for upper 50 m (full line: S; dotted line: C).

5. The variations of the surface current

As we have seen in the previous section, the surface current measured with GEK varied with time both in strong Kuroshio current and the weak current

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region. It swept clockwise and its radius was about 0.4~0.5 kt. Hence in strong current region it seems like to be steady in its velocity, but in weak current region it seems like to vary strongly. The period of its circulation did not coincide with each other. The surface current measured with GEK was affected by many factors, such as the diurnal variation of the earth magnetic field and earth electric current, wind, the tidal current and the internal wave. All the effects except that of wind were periodical and their periods resembled with each other, so that without observing it continuously we can not distinguish individuals. But it was probable that the inertia period was predominant, and this might be considered to associate with the internal wave.

6. *The variation of the dynamic depth anomaly or the geostrophic current*

The fluctuation of the isotherms varied also the specific volume distribution

Table 2. Variations of dynamic depth anomalies.

St. Observation No.	A	Sh ₂	Ry	D	T	Y	R	S	C
1	0.965	1.392	1.690	1.178	1.011	0.688	1.613	1.556	1.669
2	0.948	1.521	1.696	1.198	0.993	0.651	1.600	1.575	1.603
3	0.941	1.506	1.674	1.183	1.014	0.646	1.617	1.585	1.601
4	0.998	1.514	1.686	1.186	1.007	0.720	1.603	1.596	1.600
5	0.969	1.460	1.700	1.170	0.988	0.663	1.631	1.601	1.595
6	0.959	1.500	1.688	1.167	0.986	0.643	1.619	1.595	1.573
7	0.969	1.490	1.688	1.184	0.917	0.673	1.636	1.589	1.606
8			1.696	1.182	1.006	0.665	1.612	1.640	1.595
9			1.622	1.141	1.002	0.664	1.659	1.628	1.620
10			1.651	1.117	0.950	0.658	1.620	1.607	1.594
11				1.186		0.668	1.626	1.585	1.600
12				1.152		0.672	1.612	1.586	1.630
13				1.174			1.636	1.611	1.602
14				1.176			1.607	1.611	—
15				1.196			1.627	1.619	1.635
16				1.251			1.592	1.586	—
17				1.234			1.602	1.614	1.632
18				1.258			1.582	1.595	1.599
19							1.583	1.595	1.591
20							1.547	1.569	1.521
21							1.555	1.550	1.548
Reference level (m)	500	1000	1000	400	400	200	1200	1200	1200

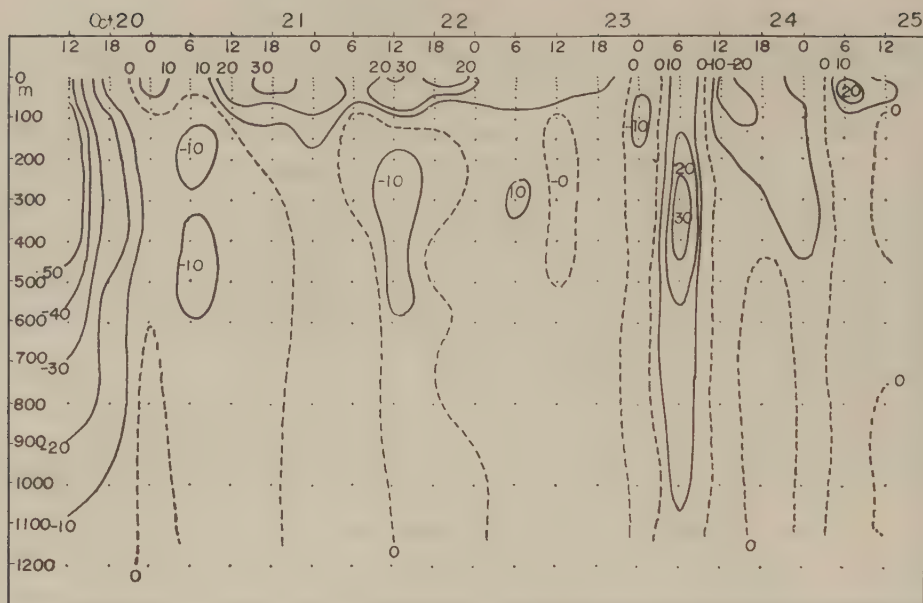


Fig. 33. Variations of geostrophic current (cm/sec.) between stations R and C (full line: R-S; broken line: C-S)

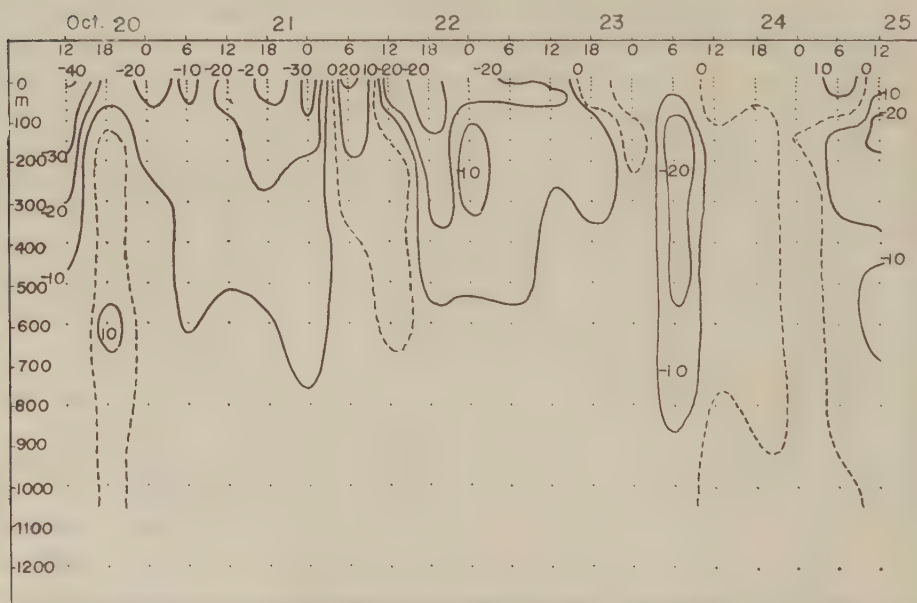


Fig. 34. Variations of geostrophic current (cm/sec.) between stations S and R (full line: R-S; broken line: C-S)

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Fig. 35. Variations of geostrophic current (cm/sec.) between stations C and S (full line: R-S; broken line: C-S)

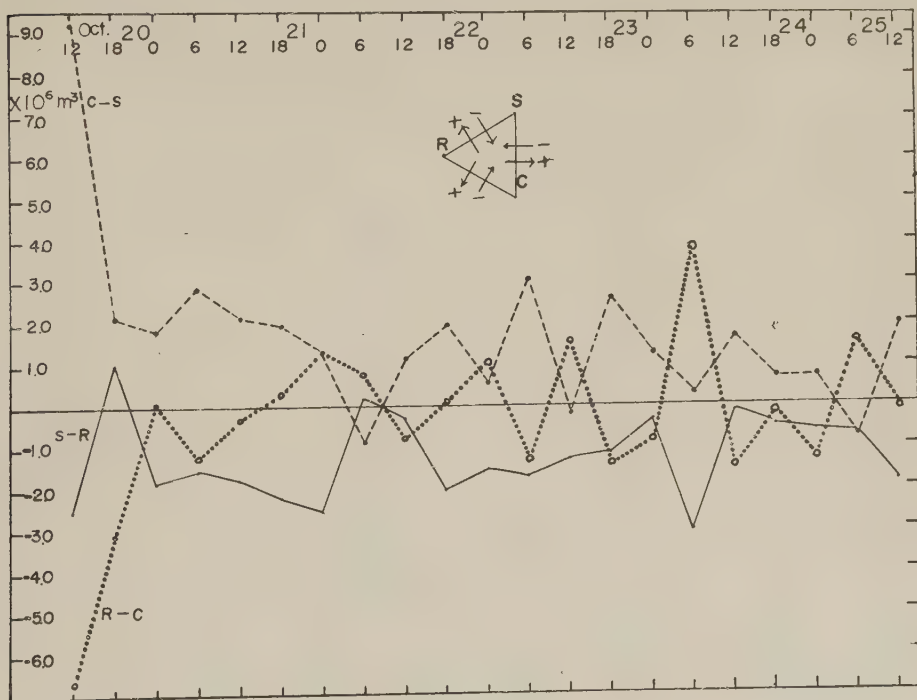


Fig. 36. Variations of mass transports ($\text{m}^3/\text{sec.}$) between the stations (full line: R-S; broken line: C-S; dotted line: R-C).

and so the dynamic depth anomalies changed. As the vertical gradient of the density below 500 m was not so large as the upper layer, the variation of the dynamic depth anomalies was perhaps mainly due to upper 500 m. In Table 2, the dynamic depth anomalies of the stations A, Ry, Sh₂, D, T, Y, R, S and C are listed. The differences were 5-14 dynamic centi-meters in one series. As the wave length of the internal wave was about 20 miles, and the difference of dynamic depth anomalies due to the internal wave was about 10 dynamic centi-meters, the corresponding change of the geostrophic current would be about 1.3 kt. The variations of the geostrophic current between the stations R, S and C are shown in Figures 33, 34 and 35. The largest variation is found between stations C and S. The maximum difference of the current was about 2.5 kt (12h Oct. 20 and 06h Oct. 23). The volume transports among them are shown in Fig. 36. The variations were rather aperiodic, but it seems that the phase was opposite between R-C and C-S. S-R was the most aperiodic. So we might suppose that the internal wave proceeded along S-R. The error of the interpolation of the water temperature expected to be about 0.5°C or less. If this error be extended to 200 m, the error of the dynamic depth anomalies would be about 2 dynamic centi-meters, and might introduce a current of about 0.3 kt. This value is not small compared with that induced by the internal wave.

7. *Vertical distribution of the current*

From the view-point of the geostrophic current, the vertical distribution of the current was quite simple in almost all the cases. The maximum value was found near the sea surface and the minimum value near the reference level. Below this level the current converted its direction. The actual current showed a very complex distribution. The current observed with the current meter contained all the motion of the water, while the geostrophic current contained only the baroclinic current. Moreover observed current was rather instantaneous value of the current. The observation was carried out at three positions along the meridian of 136°-40'E with Nan'nichi-Iwamiya current meter (100, 200, 300, 400 and 500 m) and Ekman-Merz current meter (600, 800 and 1000 m). The results are shown in Fig. 37. The reference level was decided from the shift of the vessel with loran fix, and from the surface current measurud with GEK, before and after the deep layer current measurement. Even if the reference level was false, the relative change of velocity was true. We were surprised by the fact that the

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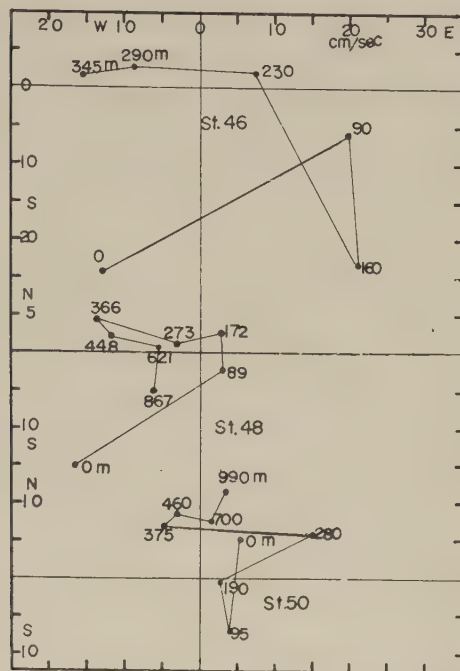


Fig. 37. Variations of the current velocity with depth. The positions of stations are as follows: st. 46 (33-05 N, 136-40 E), st. 48 (32-35 N, 136-40 E) and st. 50 (32-05 N, 136-40 E).

Table 3. Geostrophic and measured currents, both being referred to 800-db surface.

St. No. depth	46 geo. curr.	48 geo. curr.	50 geo. curr.
0	-8.8	-22.2	+18.2
10	-8.6	-22.1	+11.0
20	-8.5	-21.9	+10.9
30	-8.2	-21.0	+10.7
50	-7.7	-19.1	+10.2
75	-7.9	-18.5	+10.1
100	-8.4	-18.2	+9.9
150	-8.8	-16.2	+8.1
200	-7.8	-13.7	+6.7
300	-4.5	-8.0	+4.1
400	-2.3	-4.5	+2.1
500	-1.7	-3.6	+1.6
600	-1.4	-2.8	+1.6
800	0	0	0
depth meas. curr.		depth meas. curr.	depth meas. curr.
0	-13.0	0	+11.6
90	+19.5	90	+3.6
160	+21.0	170	+0.4
230	+7.5	275	-28.8
		365	-13.9
		450	-10.3
		620	-1.8
		800	0

current was pretty strong at deep layers where the current is usually considered as motionless, and the surface current did greatly differ from that of 100 m-layer, which was probably reduced by the internal wave. The east-west component of the measured current was compared with the geostrophic current in Table 3. We found that they were much different from each other.

8. Eddy viscosity and eddy conductivity of heat

Lateral eddy viscosity coefficient k is defined by the following equation :

$$k = -\overline{\rho u'v'} \left/ \frac{d\bar{u}}{dy} \right.$$

where u' , v' are the turbulent velocity components of x , y directions, and $\bar{u}$, the mean current, assuming that the mean current flows toward x direction. Available data were observed in the cases (c), (e), (f), (i), and (j). We observed them frequently at one position and around it. It was reasonable to consider that there was no mean current to y -direction.

Lateral eddy conductivity of heat K is to be obtained by the following equations :

$$(i) \quad K = \bar{u} \frac{dT}{dx} \left/ \frac{d^2T}{dx^2} \right. \quad (ii) \quad K' = -\overline{\rho u't'} \left/ \frac{dT}{dy} \right. .$$

These values would also be obtained for the cases (e) and (f). The computed values of k and K are listed in Table 4. The order of magnitude of these values

Table 4. Eddy viscosities (k) and eddy conductivities of heat (K and K').

Case	k	K	K'
c	-0.6×10^6		
e	-2.0×10^6	2.0×10^7	1.2×10^7
f	$+0.9 \times 10^7$	-0.9×10^7	-0.4×10^7
i	-1.5×10^6 (x) $+2.4 \times 10^6$ (y)		
j	-7×10^6 (s) $+6 \times 10^7$ (c)		

is about 10^7 , which is reasonable. But since their signs are negative, it is doubtful whether we may trust in the result. I am afraid that the lack of data or misuse of the equations may have lead to the above results. As for the former constant, to calculate the turbulence, two-hourly or four-hourly observations of the elements were available but the number of observations (the period of the observation) may

have been too small to fulfill the condition of the above equations. Especially, the current measured with GEK may have contained much uncertainties, because of the indirect method of measurement. As regards K , since the observations of the elements were not given at the same time, the variations of the values with space and time were neglected. Moreover in order to neglect the diurnal variation of the surface water temperature due to solar radiation, we used the data for layers 20 m or 100 m. The current was measured at the surface. From the discussion of the previous section, the actual currents were much different to the depth. We suppose that [these effects would make error and change the physical constant negative. If we have used the equation (ii), we have obtained the value of K' which is nearly equal to K , their signs being equal. Usually $[\frac{dU}{dy}] \cdot [\frac{dT}{dy}] > 0$ near the Kuroshio, so that it must be $\overline{u't'} \cdot \overline{u'v'} < 0$. This relation suggests that t' and v' were of opposite sign. This relation is rather contrary to the general distribution.

9. Concluding remarks

For future work, the observation of the internal waves from the drifting vessel should be carried out carefully; her position should be kept within 1 mile from the fixed point, because the speed of the internal waves is about 1.5 kt and its wave length is about 20 miles. The vessel had better be anchored, though it is very difficult in deep ocean. As inertia and tidal periods resemble to each other in the sea near Japan, it is difficult to distinguish between them. A greater difficulty is the generation and decay of the inertia wave by irregular meteorological disturbances. From the observation made at an anchored station R in case (j), the period of an internal wave was known to be about 11.7 hours. This is expected to have occurred from the superposition of M_2 tide and the wave with half period of the inertia wave. The component waves with such a small difference of period as this can not be distinguished from each other by these short-period observations. We may say that the observed internal wave was not a single wave (tide or inertia), but a superposed wave. The inertia wave is generated by current, wind etc., and the direction of the internal wave distributes rather randomly. There are so many unknown factors of the internal wave that it is not easy to catch the general character of the wave.

We can conclude that for more accurate observations of the internal wave, three or more vessels anchored about 5 miles apart from each other are to be

used. Though BT is available for the study of internal tide or inertia wave, it is likely to use internal waves of shorter period may disturb the records. So it is desirable to use a more developed equipment, such as, an isotherm follower or other multi-continuous temperature recorder.

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